Engineering Hydrology

B.E. (Agricultural Engineering & Civil Engineering)

Lecture notes

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By IOE Agrineer

Chapter 1: Introduction

1.1 General introduction

Hydrology is derived from two words: hydro and logos. 'Hydro' means water and 'logos' means study. Simply, Hydrology is defined as the study of water.

Hydrology is the science which deals with the origin, distribution and circulation of water in different forms in land phases and atmosphere.

Broad definition of Hydrology given by US National Research Council

Hydrology is the science that treats the waters of the Earth, their occurrence, circulation and distribution, their chemical and physical properties, and their reaction with the environment, including their relations to living things.

Interdisciplinary subject: As an earth science, Hydrology is connected to several subjects. These include:

- Meteorology: for understanding precipitation and evaporation process
- Soil science: for understanding infiltration
- Geology: for understanding groundwater flow
- Geomorphology: for understanding surface runoff
- Hydraulics: for understanding stream flow

  Physics, Chemistry, Biology, Math, statistics: to formulate and understand the subject

Division of hydrology

1. Scientific hydrology: deals with hydrological processes from the view point of natural processes
2. Applied or engineering hydrology: deals with engineering applications of hydrology

Scope of hydrology

1. Estimation of water resources
2. Study of processes like precipitation, evaporation, infiltration and runoff and their interaction
3. Study of problems like floods and droughts
4. Understanding the properties of water in nature

Things to be considered for planning and design of water resources projects

a. Maximum flows which are expected to occur at a place
b. Minimum flows which can occur during any dry period
c. Minimum reservoir capacity to be fixed to meet all water demands from a multipurpose reservoir
d. Possible regulation of floods at the downstream reaches once a hydraulic structure is erected
e. Possible supply of water from a river to meet demands for water resources projects
f. Environmental impacts of a hydraulic structure
g. Study of groundwater potential and its use

Watershed/catchment/drainage basin

Watershed is the area of land draining into a stream at a given location. Divide is a line which separates catchment from its neighbouring catchments. For delineating basin, we need topographic map. The map shows changes in elevation by using contour lines.

Features of contour

- Uphill: contour with higher elevation
- Hill: circular contour, ridge: highest point
- Saddle: mountain pass
- Valley: V or U shaped with the point of the V/U being the upstream end
- Close together contours: steep slope
- Widely spaced contour: level ground

Basin delineation procedure on topo map

- Mark the outlet point
- Mark the highest point around the river
- Start from the outlet and draw line perpendicular to the contours in such a way that the line passes from the highest point (ridge)
- Continue to the opposite side of the watercourse, finally ending to the outlet.

Finding area of watershed/basin

Method 1: Use planimeter around the boundary

Method 2: Trace the basin and count area manually

Method 3: Use of GIS package

1.2 Applications of Hydrology in Engineering

- Correct assessment of flows for hydropower, irrigation, drainage and water supply projects.
- Determination of maximum expected flow at dam, reservoir, spillway, bridges, culverts and city drainage system.
- Determination of minimum reservoir capacity sufficient to meet the hydropower, irrigation and water supply demands.
- Estimation of the total volume of water that may be available from a drainage basin over a long period
- Flood control: statistical analysis of probable frequency of floods, estimation of design flood, and flood forecasting.
- computation of water surface profile for various rates of flow for navigation
- Control of erosion to minimize sedimentation of reservoirs.
- Reduction of stream pollution

1.3 Hydrological cycle

The endless circulation of water between the earth and its atmosphere is called hydrological cycle. Hydrological cycle is the most fundamental principle of hydrology. The cycle extends its scope from 15 km up into the atmosphere from the earth's surface to about 1km below the earth's crust through a maze of paths. It is fueled by solar energy and driven by gravity force.

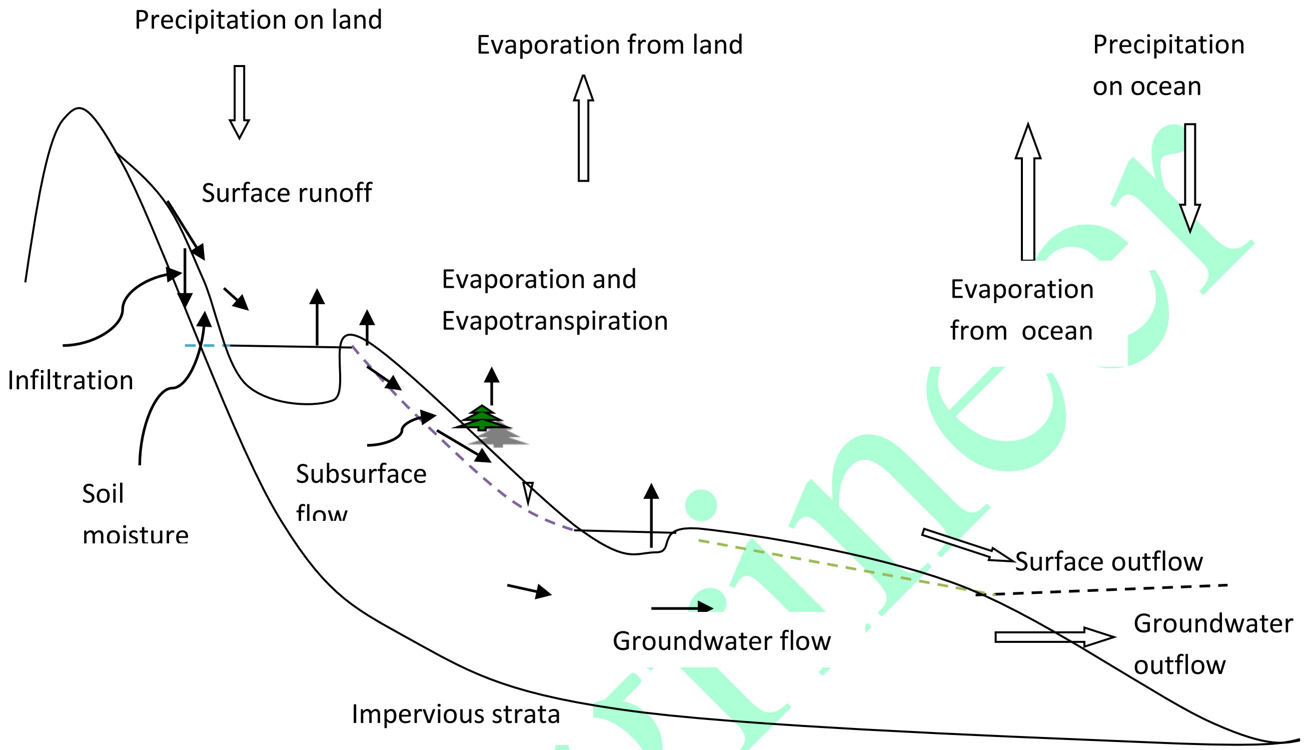


Fig. 1.1: Hydrological cycle

Processes

a) Evaporation: Water is evaporated from the oceans and land surfaces to become part of the atmosphere.
b) Precipitation: Water vapour is transported and lifted in the atmosphere until it condenses and precipitates (falls in the form of solid or liquid) on the land or the oceans.
c) Interception: Part of precipitation is intercepted by vegetation and trees.
d) Infiltration: Part of precipitation infiltrates into the soil.
e) Surface runoff (Overland flow): The fallen precipitation flows over the land surface before reaching the channel
f) Evaporation and Transpiration: Much of the intercepted water and surface runoff returns to the atmosphere through evaporation. Part of the infiltrated water is available to the roots of the trees and returns to the atmosphere through plant leaves by transpiration.
g) Subsurface runoff (Interflow): The infiltrated water flows laterally through the unsaturated soil to the stream channel.
h) Deep percolation: The water from the soil moisture zone percolates deeper to recharge ground water.
i) Ground water flow (Base flow): The flow takes place from the saturated groundwater zone to the streams.
j) Final output: Streamflow
   - The part of precipitation that reaches the stream through different paths above and below the earth surface is called runoff. Once it enters the channel, the runoff is called streamflow.
k) Finally the precipitated water flows out into the sea which it will eventually evaporate once again and the hydrological cycle continues.

1.4 Water budget or water balance equation

The water balance equation is the statement of the law of conservation of mass. Water balance is the balance of input and output of water within a given area taking into account net changes of storage.

Change in storage = Inflows-Outflows

$$\frac{d}{dt}(Storage) = Inflows - Outflows$$

It is also called continuity equation or conservation equation.

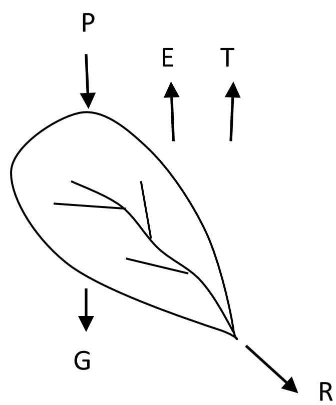


Fig. 1.2: Various components of water balance in a basin

General water budget equation in hydrology for time interval Δt

$$P - (R + G + E + T) = \Delta S$$

P= precipitation
R = Surface runoff
G = Net groundwater flow out of the catchment
E = Evaporation
T = Transpiration
ΔS = change in storage (take + for increase in storage, and – for decrease in storage)

All the terms in the equation have the dimensions of either volume or depth.
Conversion of unit given basin area: Volume = Depth x basin area
Conversion to volume given flow rate: Volume = flow rate x time duration

In case of other inflow besides precipitation, the water balance equation is
$(P + I) - (R + G + E + T) = \Delta S$ where I= other inflow

For long term, e.g. annual water balance, change in storage is zero. The general water balance equation is:

Precipitation-Runoff=Evaporation

Significance of water balance equation
The water balance equation is useful to assess the various components of the hydrological processes for a certain time interval. The input to the system (basin) is precipitation, and the outputs from the system are surface runoff, evaporation, transpiration and groundwater flow. Storage component is the water stored within the basin. The assessment of runoff using water balance equation is useful for water resources

projects, such as irrigation, water supply, flood control, pollution control etc. The equation is also useful for estimating the change in storage in a reservoir and estimating losses from precipitation.

1.5 History of hydrological development in Nepal

History of hydrological development in Nepal is not very long. Different activities in the development of hydrology of Nepal are summarized below.

a. Preliminary works in the period of 1940-1960

- Starting of Hydrological studies in Nepal after the Government of India initiated Koshi project in late 1940s.
- Establishment of Hydrological stations on Koshi at Barahachhetra, Sunkoshi at Kampughat and Tamur at Mulghat in 1947.
- Establishment of meteorological observations stations in 1956 with the support of the Government of India.

b. Establishment of Department of Hydrology and Meteorology

- Nepal started hydrological and meteorological activities in an organized way in 1962 from the Karnali basin. The activities were initiated as a section under the Department of Electricity.
- Establishment of the Department of Hydrology and Meteorology (DHM) under the ministry of Water and Power in 1966.
- Publication of hydro-meteorological data from 1966.
- Merging of the DHM with the Department of Irrigation in 1972.
- Separation of DHM in 1988 from the irrigation.
    - Main responsibilities of DHM: collection, analysis, processing, dissemination of hydrometrorological data; meteorological and hydrological forecasting; research work on hydrology and meteorology

- Starting of Nationwide hydro-meteorolgical data management project in 1993.
- From recent years, use of modern technology for data collection such as wireless communication, satellite data receiving system, receiving data through internet using CDMA

c. Station network

Koshi, Gandaki, Karnali, Mahakali, Bagmati, Kamala, Kankai, Babai, West Rapti are major basins of Nepal. The topography of Nepal has a major role in determining the hydrological network. The hydrological network in Nepal is very poor in headwater region, fair in mountainous region and again poor in the Terai region. At present, DHM maintains 154 hydrological stations and 337 precipitation stations all over Nepal.

d. Acts, plan

I. Water Resource Act (1992): act enacted for rational utilization, conservation, management and development of the water resources of Nepal

II. Water Resources Strategy (WRS, 2001): strategies formulated to improve the living standard of people through the water resource development

III. National water plan (NWP, 2005): In order to implement the activities identified by the WRS, the Water and Energy Commission Secretariat (WECS) formulated National Water Plan (NWP) in 2002, which was approved in 2005. The NWP is a framework to guide, in an integrated and comprehensive manner, all stakeholders for developing and managing water resources and water services.

Chapter 2: Precipitation

2.1 Precipitation

Precipitation is any form of solid or liquid water that falls from the atmosphere to the Earth's surface.

Forms of precipitation

- Drizzle: water droplets, low intensity, 0.1 – 0.5 mm
- Rain: water droplets, higher intensity, > 0.5 mm
- Glaze: ice coatings formed by freezing rain
- Sleet: ice grains formed due to freezing temperature while falling
- Snow: ice crystals, hexagonal
- Hail: ellipsoid ice balls, 5 to 125 mm
- Dew: during clear nights, when the surface of the object on the earth cools due to radiation, the moisture present in the atmosphere condenses on the surface of these objects forming water droplets called dew.

Requirements for precipitation formation
1. Uplift of air mass into atmosphere
2. A gradient in temperature in the atmospheric column
- Decrease of air temperature with altitude influences the amount of moisture that can be held as water vapor in air.
3. Water vapor in the atmosphere and saturation as temperature changes
In higher parts of the atmosphere (colder), there is less ability to hold water vapor (decrease in vapor pressure).
– Implies that colder air holds less water vapor, leads to saturated conditions (moist air) and promotes condensation.
4. Presence of nuclei (salt, dust, and clay around 1 to 10 μm in diameter, called aerosols) around which condensation of vapor takes place.
5. Precipitation product must reach the ground in some form.

Mechanism of precipitation formation
- Water vapor rises in the atmosphere and cools.
- Water droplets in clouds are formed by nucleation of vapor on aerosols.
- Droplets increase in size by condensation.
- Droplets (~0.1mm) become heavy enough to fall.
- Many of the falling droplets decrease in size by evaporation, which are again carried upwards in the cloud.
- Some of the falling droplets increase in size by impact and aggregation. Some of the larger drops (3-5mm) may break up into smaller raindrops and droplets. The droplets may be again carried upwards in cloud.
- When the diameter of droplets becomes 0.1-3mm, they start falling.

Front
A front is the interface between two distinct air masses.

Types of precipitation based on lifting mechanism

1. Convective

Unequal heating at the surface of the earth is the main cause of convection. In summer days air in contact with the surface of the earth gets heated up, expands and rises due to lesser density. Surrounding cold air rushes to replace it and in turn gets heated up and rises thus setting up a convective cell. The warm air continues to rise and undergoes condensation. The condensation releases latent heat of vaporization, which helps to move the air mass up. Depending on the moisture content, cooling and other factors, the precipitation intensity varies from light showers to cloud bursts. Sometimes upward wind currents exceeding 150 kmph freezes the raindrops to form hail.

2. Orographic

Lifting of air mass over a mountain barrier is called orographic lifting. Dynamic cooling takes place causing heavy precipitation on the windward side and light on leeward side. Orographic precipitation gives medium to high intensity rainfall and continues for longer duration.

3. Cyclonic

A cyclone is a low pressure region surrounded by a larger high pressure area. The cyclone center is called eye, which is a calm area. This zone is surrounded by strong wind zone. The pressure decreases towards eye.

When the low pressure occurs in an area, especially over large water bodies, air from the surroundings rushes, causing the air at low pressure zone to lift. The system derives its energy from sea vapor. Once the cyclone crosses over to the land, the energy source is cutoff, it becomes weak and disappears quickly. The rainfall is normally heavy in the entire zone travelled by a cyclone.

An anticyclone is an area of high pressure in which wind tends to blow spirally outward in clockwise direction in the northern hemisphere and anticlockwise in the southern hemisphere. Weather is usually calm and such anticyclones are not associated with rain.

2.2 Rainfall measurement

Precipitation is measured as depth of water equivalent from all forms that would accumulate on a horizontal surface if there are no losses.
Unit: mm or inch

Methods of precipitation measurement

- Rain gauge
- Radar
- Satellite

Types of rain gauge
1. Non-recording gauge
The gauge which is read manually is called non-recording gauge. It does not record rain itself, but simply collects. It consists of collector above funnel leading into receiving vessel. The rainfall collected in the vessel is measured by a graduated measuring cylinder or dipstick to give depth of rainfall.

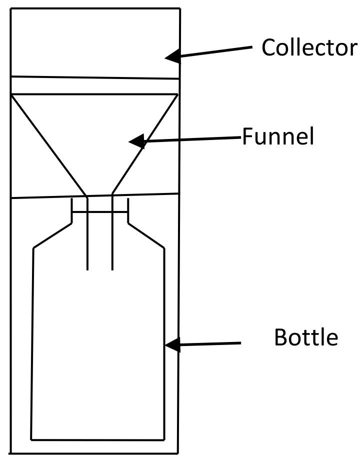


Fig. 2.1: Non-recording gauge

2. Recording gauge

The gauge which records the depth of rainfall automatically is called recording gauge. Rainfall intensity, duration and depth can easily be obtained from recording gauge. There are three types of recording gauge in general use.

a) Tipping bucket: Tipping bucket type gauge operates with a pair of buckets. When the rainfall first fills one bucket, it tips and brings the other one in position. The flip-flop motion of the tipping buckets is transmitted to the recording device (clock-driven drum chart) and provides a measure of rainfall intensity. Alternatively, the tipping mechanism is used to actuate electric circuit which records the number of tips during rain. Usually one tipping is equal to 0.25mm of rain. The instrument is suitable for digital data.

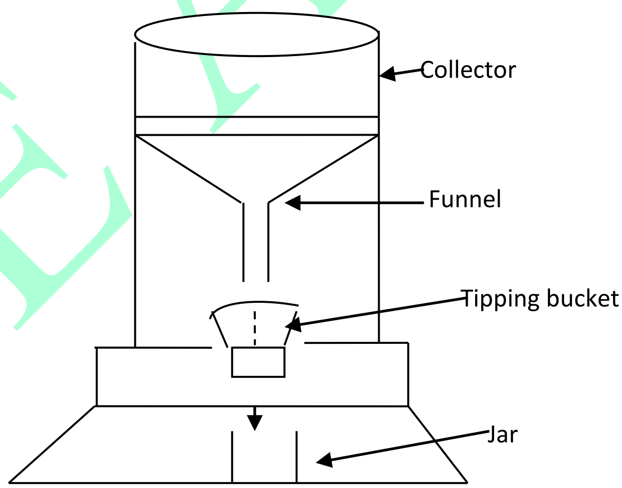


Fig. 2.2: Tipping bucket rain gauge

b) Weighing bucket: In this gauge, rainfall is collected in bucket which rests on a weighing scale with a spring mechanism. For recording the rainfall, mechanical lever arm of the balance is connected with a pen which touches a clock mounted drum with a graph paper. The record shows accumulation of rainfall over time.

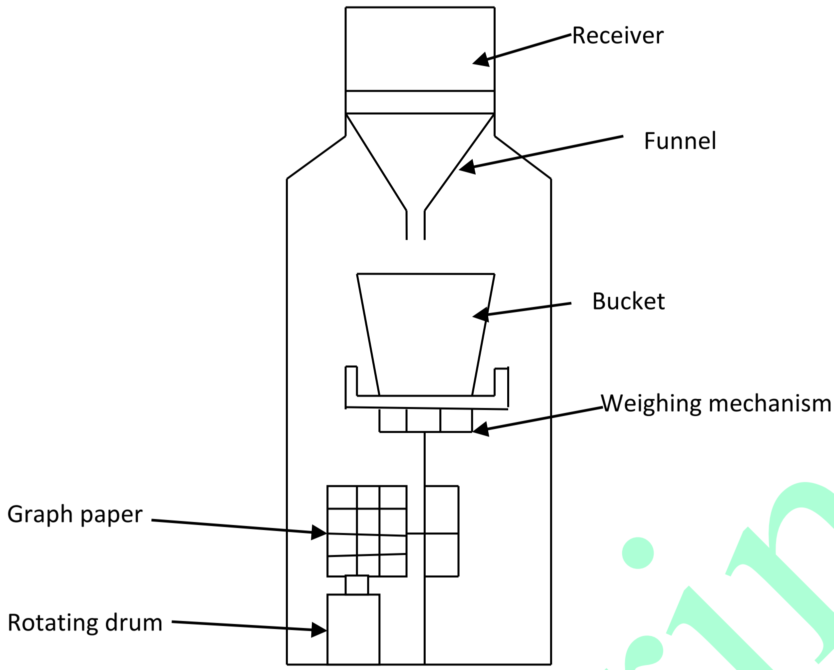


Fig. 2.3: Weighing bucket rain gauge

c) Float type (Syphon) gauge: This type of gauge has a chamber containing a float. With the increase in rainwater in the chamber, the float rises. Vertical movement of the float is translated into movement of a pen on a chart, which is mounted on a mechanical clock. A syphon arrangement empties the float chamber when the float has reached the pre-set maximum level. Then the pen comes back to original zero position showing vertical line on the graph. If there is no rainfall, the pen moves horizontally. Each syphonic action measures certain amount of rainfall, e.g. 10mm. This instrument provides mass curve of rainfall.

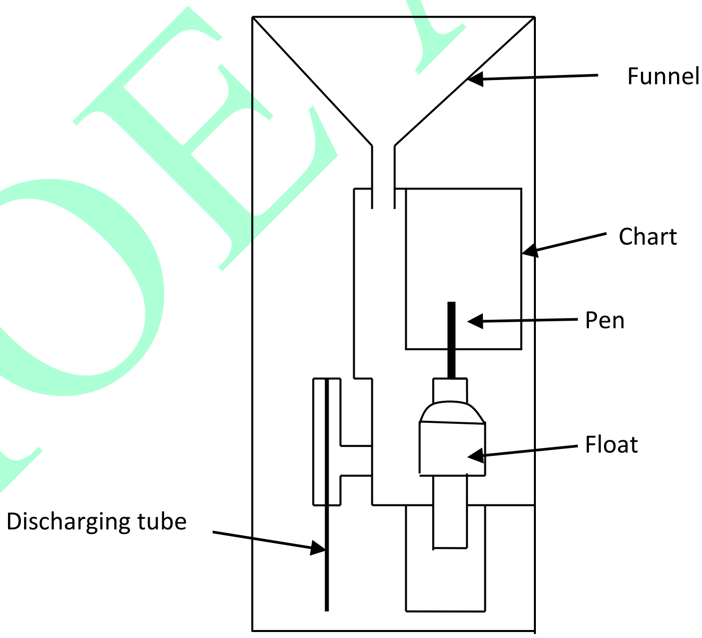


Fig. 2.4: Float type rain gauge

Error in measurement

- Instrumental error
- Human error
- Wind error
- Evaporation error
- Wetting error
- Splashing error

Telemetry

Telemetry is a system of transmitting rainfall data to a base station by means of electronic units connected to recording gauges. This system is very useful for mountainous and inaccessible areas. Telemetry is used for real time transmission of data.

2.3 Design of rain gauge network

WMO recommendations

Type of regions	Minimum area for one station under ideal condition in sq.km.	Area to be covered under difficult condition per station in sq.km.
1. Flat regions of temperate and Mediterranean and tropical zones	600-900	900-3000
2. Mountainous regions of temperate and Mediterranean and tropical zones	100-250	250-1000
3. Small mountainous regions with irregular precipitation	25	
4. Arid and polar zones	1500-10000	

Optimum number of raingauge stations

Records from all the existing gauges of a basin help to fix the optimum number of stations. The following statistical analysis helps to obtain optimum number of gauges for a basin on the basis of an assigned percentage of error in estimating the mean areal rainfall.

$$N = \left(\frac{C_v}{E_p}\right)^2$$

N = optimal number of stations
E_p = allowable percentage of error in estimating the mean areal rainfall
C_v = Coefficient of variation of the rainfall from existing stations
Method to calculate coefficient of variation
Mean of rainfall: $P_{av} = \frac{1}{n}\sum P_i$

Standard deviation: $\sigma = \sqrt{\frac{1}{n-1}\sum(P_i - P_{av})^2}$

Coefficient of variation: $C_v = \frac{\sigma}{P_{av}} \times 100$
Ep is generally taken as 10%.
If Cv is less than 10%, the existing number of stations is assumed to be sufficient. For N>n, the additional stations required for the basin are N-n. Annual rainfall data is normally used in this analysis.

Normal rainfall
Average rainfall for 30 year period

2.4 Estimation of missing precipitation

Two commonly used methods
1. Arithmetic average method

This method is used if the normal annual rainfall of missing station is within 10% of the normal annual rainfall of surrounding stations, data of at least 3 surrounding stations (index stations) are available and the index stations should be evenly spaced around missing station and should be as close as possible.

The formula for computing rainfall of missing station is

$$P_x = \frac{1}{n}(P_1 + P_2 + \cdots + P_n)$$

$P_1, P_2, \ldots, P_n$: rainfall of index stations
P_x: rainfall of missing station
n: number of index stations

2. Normal ratio method

This method is used if the normal annual rainfall of index stations differs by more than 10% of the missing station. The rainfall of surrounding index stations is weighed by the ratio of normal annual rainfall by using the following equation:

$$P_x = \frac{1}{n}\left(\frac{N_x}{N_1}P_1 + \frac{N_x}{N_2}P_2 + \cdots + \frac{N_x}{N_n}P_n\right) = \frac{N_x}{n}\left(\frac{P_1}{N_1} + \frac{P_2}{N_2} + \cdots + \frac{P_n}{N_n}\right) =$$

$P_1, P_2, \ldots, P_n$: rainfall of index stations
P_x: rainfall of missing station
n: number of index stations
N_x: normal annual rainfall of missing station
$N_1, N_2, \ldots, N_n$: normal annual rainfall of index stations

2.5 Double mass curve analysis for correction for data inconsistencies

The plot of accumulated annual rainfall of a particular station versus the accumulated annual values of mean rainfall of surrounding stations is called double mass curve. This technique is used to check the consistency of rainfall data and to correct erroneous rainfall data. This technique is based on the principle that a group of sample data drawn from its population will be the same.
The reasons for inconsistency are:

- Shifting of gauge
- Change in site conditions due to calamities, e.g. fires, landslide
- Change in observational procedure
- Observation error

If the double mass curve is straight line, the rainfall of the particular station is said to be consistent. If there is break in the slope of the plot, then the rainfall of that particular station is inconsistent. Starting year of change of regime of rainfall is marked by the starting point of the break in slope. Correction has to be applied beyond the period of change of regime.

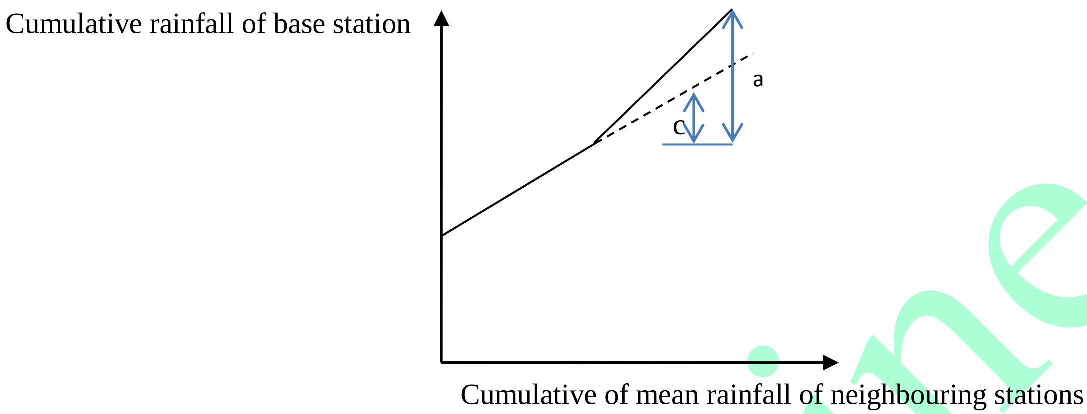


Fig. 2.5: Double mass curve

Steps in double mass construction:

- Select a group of 5 to 10 neighbouring stations
- Arrange data in chronological order with latest data in the beginning.
- Compute cumulative rainfall of base station (station whose consistency is to be checked).
- Compute cumulative of mean rainfall of neighbouring stations.
- Plot cumulative rainfall of base station versus cumulative of mean rainfall of neighbouring stations, and join the points by straight line.
- Check if there is break in straight line.

Formula for correction for rainfall after break in line is given by

$$P_{cx} = P_x \frac{M_c}{M_a}$$

P_{cx} = Corrected precipitation for station x
P_x: Original precipitation of station x
M_c: Slope of original line
M_a: Slope of line after change of regime
If c and a are vertical intercept of original line and line after change

$$P_{cx} = P_x \frac{c}{a}$$

A change in slope is normally taken as significant only where it persists for more than five years. Correction should be applied for change in slope exceeding 10% of original line.

2.6 Presentation of rainfall data

a. Rainfall depth

- Rainfall intensity: rainfall depth/time interval

b. Point rainfall

- Rainfall data of a station of certain duration
- Duration: Hourly, Daily, Weekly, Monthly, Seasonal, annual
- Plot: Rainfall versus time in bar diagram

c. Moving average

- Average of consecutive interval
- Interval: 3-5 year
- Purpose: to isolate the trend in rainfall data and to smoothen out the high frequency fluctuations

d. Mass curve

- Plot of accumulated rainfall versus time
- Useful to identify intensity, duration, magnitude, starting and ending time of rainfall
- Magnitude = cumulative rainfall at t- cumulative rainfall at t-1
- Intensity = slope of curve

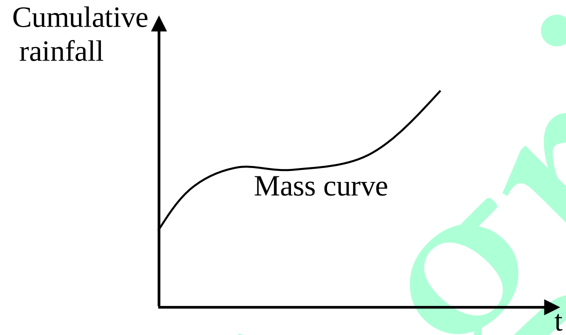


Fig. 3.6: Mass curve of rainfall

3. Hyetograph

- Plot of rainfall intensity or rainfall depth versus time interval in the form of bar graph
- In each bar, time interval between two points is shown in X-axis and corresponding rainfall represents Y-axis.
- From mass curve, rainfall of certain interval dt can be computed and rainfall intensity can be obtained.
- The graph represents the characteristics of storms and useful in predicting floods.
- Area under hyetograph: total rainfall

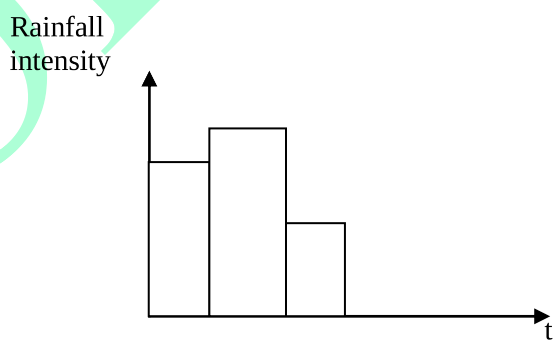


Fig. 3.7: Hyetograph

2.7 Method of computing average rainfall

Three common methods
1. Arithmetic mean method

This is the simplest method for computing mean rainfall. This method is satisfactory if the gauges are uniformly distributed and the individual gauge catches do not vary greatly about the mean. The formula for computing mean rainfall is

$$P_{av} = \frac{1}{n}\sum_{i=1}^{n} P_i$$

P_{av} = average rainfall
n = number of stations
P_i = precipitation of station i

This method gives equal weights to each gauge. It gives only rough estimate. It does not take into account the topography and other influences. For this method, only the gauges inside the basin are considered.

2. Thiessen polygon method

In this method, weightage is given to all the gauges on the basis of their areal coverage. Thiessen method assumes that rainfall at any point within the polygon is same as that of the nearest gauge. For this method, all the gauges in and around the basin are considered.

Method of constructing Thiessen polygon

- Draw map of basin and locate the rainfall stations.
- Connect the adjacent rainfall stations by straight lines, forming triangles.
- Draw perpendicular bisectors to each of the sides of triangles lines. The perpendicular bisectors forms boundary of polygons. Wherever the basin boundary cuts the bisectors, it is taken as the outer limit of the polygon.
- Measure area of each polygon which surrounds a station. Area can be computed by using formulae for regular geometric figure. For irregular figure, area is determined by using planimeter or by tracing on graph and counting squares.

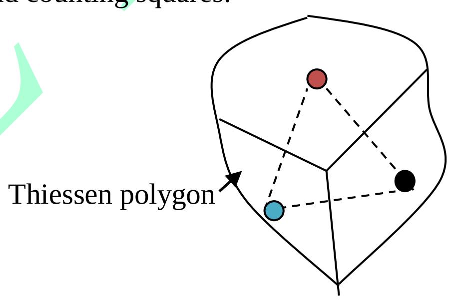


Fig. 3.8: Example of Thiessen polygon

The formula to compute mean rainfall is
$P_{av} = \frac{\sum_{i=1}^{n} A_i P_i}{A}$
P_{av} = average rainfall
P_i = Rainfall of station i
A_i = Area of polygon which encloses station i

A= Total basin area
A_i/A: Weight for station i = w_i
$P_{av} = \sum_{i=1}^{n} w_i P_i$

Advantages

- Use of data nearby stations located outside basin
- Consideration of spacing of stations
- Easy to perform computation through computer software

Limitations

- It does not consider orographic and topographic effects.
- The method assumes linear variation of precipitation between stations.

3. Isohyetal method

An isohyet is a line joining points of equal rainfall. For this method, rainfall stations lying within basin as well as nearby stations around the basin are considered.

Method of constructing isohyetes

- Draw map of basin and locate the rainfall stations.
- Mark the depth of rainfall at each station.
- Draw isohyetes by interpolating between adjacent gauges and considering orographic, storm characteristics and other factors.
- Measure the area between successive isohyets.

The mean rainfall is computed by

$$P_{av} = \sum \frac{A_j \dfrac{P_i + P_{i+1}}{2}}{A}$$

P_{av} = average rainfall
Pi, Pi+1: rainfall of isohyets i and i+1
A_j = Area enclosed by isohyets i and i+1
A= Total basin area
Note: if the isohyets go out of the boundary, then the catchment boundary is used as the bounding line.

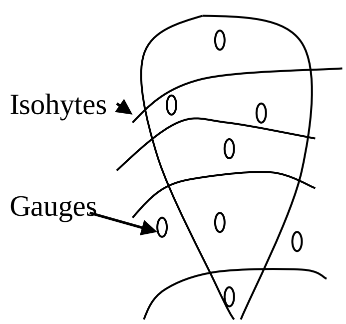


Fig. 3.9: Example of isohyetal method

Advantages

- Data from nearby stations located outside the basin can also be used.
- Spacing of station as well as magnitude of precipitation is considered in the method.
- The method is more accurate due to the consideration of topography and other influences.

Limitations

- The method requires dense gauge network.
- Isohyets need to be drawn for each storm.

2.8 Intensity duration frequency (IDF) curve

An intensity duration frequency (IDF) curve is a three parameter curve in which duration is taken on x-axis, intensity on y-axis and the return period or frequency as the third parameter. The IDF curve is a very important tool for determination of runoff, which is required for design. The curve can be used to determine the rainfall intensity for other durations with given intensity of a particular duration.

IDF Curves by frequency analysis

When observed rainfall data of different durations are available, IDF curves can be developed using frequency analysis.

Method

For each duration selected e.g. 15 min, 30 min, 1hr etc., extract annual maximum rainfall data from historical record.

For each duration, perform frequency analysis i.e. find rainfall value of different return periods such as 2, 5, 10, 25, 50, 100 yr. (Return period is the average interval of time within which an event of given magnitude will be equaled or exceeded.)

The steps involved in simple plotting position method, which can be used for precipitation data, are given below.

- Prepare data of maximum intensity for different durations for different years.
- Arrange data in descending order.
- Assign rank of data. Assign 1 for highest data, 2 for second highest data and so on.
- Calculate the return period of each data. According to California formula, return period (T) = n/m where n = number of data, m = rank.
- Plot rainfall versus return period and extrapolate to get rainfall of higher return periods.

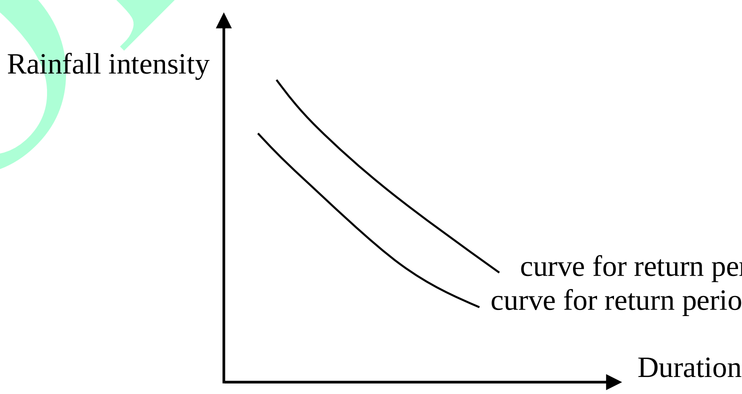


Fig. 2.10: IDF curve

Intensity of rainfall decreases with the increase in duration of storm and increases with the increase in frequency of storm. IDF curves can be expressed as equations in the exponential form given by

$i = \frac{KT^x}{(D+a)^n}$

i= intensity
T = return period or frequency
D = Duration
K, x, a, n = Constants

2.9 Depth Area Duration (DAD) Curve

A Depth-Area-Duration (DAD) curve is a graphical representation of depth of precipitation and area of its coverage with duration of occurrence of storm as third parameter. Storms of smaller duration has smaller depth and the depth decreases with increase in area.

The purpose of DAD curve is to determine the maximum amount of precipitation that have occurred over various sizes of drainage area during the passage of storm periods of say 6hr, 12hr, 24hr or other durations. Such information is essential for the design of hydraulic structures such as reservoir, dam, culverts etc.

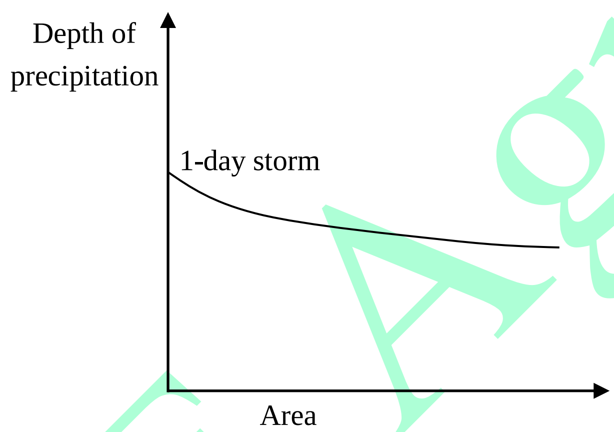


Fig. 2.11: DAD curve

Incremental isohyetal method is generally used to construct DAD curve. The procedure of this method is as follows.

Identify all the major storms.

Note down the duration for all storms, such as 1 day duration.

Prepare isohyetal pattern for all 1 day storms on map.

Take one 1 day storm and calculate the area bounded within the highest isohyets by planimetering. Next compute the area bounded between the largest and second largest isohyets by planimetering. Repeat the procedure for remaining isohyets. Compute average isohyetal depth (P_{mi}). This procedure is repeated for all other 1 day storm of the area. The average depth of precipitation is computed as follows.

$$d_n = \frac{\sum_{i=1}^{n} P_{mi} A_i}{\sum_{i=1}^{n} A_i}$$

where d_n = average depth of precipitation covering up to nth isohyets, P_{mi} = average isohyetal depth for i^{th} isohyets, A_i = Area enclosed between i^{th} and $(i+1)^{th}$ isoheyts.

Plot a graph between cumulative area as abscissa and maximum average depth of precipitation as ordinate covering the depth area data of all 1 day storms.

Same procedure as explained above is applied for other durations.

DAD curve can be expressed by empirical equation as

$$P = P_0 e^{-KA^n}$$

P= depth of rainfall over an area of A, P_0 = rainfall at center of rainfall, k and n = constants

2.10 Snowfall and its measurement

The atmospheric requirements for snow fall

- Presence of water vapor
- Presence of ice nuclei
- Ambient temperature below 0^0C

Ice nuclei: particles that cause ice crystals to form through either direct freezing of cloud droplets or freezing of water deposited on the particle surface as vapor. E.g., dust particles, combustion products, organic matter

Once ice crystals form, they may splinter and create large number of nuclei to aid the precipitation process. Continued growth of an ice crystal leads to the formation of a snow crystal.
Snow pellets
Snowflake: aggregation of snow crystals, may grow in size during its falling

Snowfall or rainfall from snowflake depends upon extent and temperature of layers of air through which it falls.

Variables

- Depth of snow
- Snow water equivalent
  Snow water equivalent is the amount of water that would be obtained if the snow were melted.
- Density of snow
  It is the percentage of snow volume that would be occupied by its water equivalent.
  Density = volume of melt water from a snow sample/intial volume of sample

Snow water equivalent = Depth of snow x Density of snow
Density: in terms of fraction
If density of water and density of snow is given,
Snow water equivalent = Depth of snow x (Density of snow/Density of water)

Density of new snow: 0.01-0.15, damp new snow: 0.1-0.2, settled snow: 0.2-0.3

Measurement of snow

1. Point snowfall water equivalent: Snow is collected in raingauge (non-recording, weighing type), which is melted and equivalent amount of water is measured.
2. Point snowfall depth: Ruler, snowboard
Snow accumulated on the snow board is measured by scale or ruler
3. Snow depth for area having large accumulation: permanent snow stakes (calibrated wooden posts fixed on the ground)
3. Areal snow cover depth and density: Snow survey
Snow survey means surveying of sections of snow cover to determine depth and density.
Depth determination: by preinstalled gauges
Density determination: boring a hole through the snowpack or into the pack and measuring the amount of liquid water obtained from the sample.

Chapter 3: Hydrological Losses

3.1 Different losses

The difference between precipitation and runoff can be treated as hydrological losses.

a. Initial losses (interception and depression storage)

Interception
Interception is that part of precipitation which is caught and held by the vegetation or obstruction. Much of the intercepted water returns to the atmosphere by evaporation. The remaining part may drip off or flow down through the stem to reach the ground surface. About 10 to 20% of total rainfall is considered as interception losses. Its exact estimation is difficult.

Depression storage
After precipitation of a storm reaches the ground, some part of it is stored in the depressions on the ground surface, which is called depression storage. The amount is eventually lost to runoff through process of infiltration and evaporation and thus forms a part of the initial loss.
The depression storage depends upon

- The type of soil
- The condition of the surface reflecting the amount and nature of the depression.
- The slope of the catchment
- The antecedent precipitation, as a measure of soil moisture.

b. Evaporation
The process by which liquid is converted to vapor is called evaporation. Evaporation occurs from water bodies as well as from soil moisture.

c. Transpiration
The emission of water vapour from plant leaves is called transpiration.

d. Infiltration
Infiltration is the process by which water from the ground surface enters into the soil. Infiltration is responsible for recharging groundwater and for maintaining soil moisture.

3.2 Evaporation process

3.2.1 Meteorological parameters

a. Temperature

Temperature is a measure of hotness of an object. The temperature of a locality is a complex function of several variables such as latitude, altitude, ocean currents, distance from sea, winds, cloud cover, and aspect (land slope and its orientation).

Lapse rate

The rate at which temperature decreases with increase in altitude is called lapse rate. It is about 6˚C per 1000 m within the troposphere.

Terminologies for expressing temperature

- Mean daily temperature: Average of hourly temperature, if hourly data are available
- Maximum daily and minimum daily
- Average of the daily max and min temperature, if only maximum and minimum data are available
- Normal temperature: Arithmetic mean temperature based on previous 30 years' data
- Normal daily temperature: The average mean daily temperature of a given date computed for a specific 30-year period.
- Mean monthly temperature: average of the mean monthly maximum and minimum temperature.
- Mean annual temperature: average of the monthly means for the year.

Temperature measurement

- Using thermometer
- The maximum-minimum thermometers for daily maximum and minimum temperature.

b. Humidity

Amount of water vapor in air is called humidity. Humidity is closely related to its temperature- higher the air temperature, more vapor the air can hold. For this reason, saturation vapor pressure goes up with air temperature.

Saturation vapor pressure

Pressure at which air is saturated with water is called saturation vapor pressure. It is a function of temperature.

$$e_s = 611\, exp\left(\frac{17.27T}{237.3 + T}\right)$$

e_s = saturation vapor pressure (N/m^2)

T= Temperature (^0C)

Significance of Humidity: The amount of water vapor in air effectively controls the weather condition by controlling evaporation from land and water surfaces.

Commonly used measures of humidity

I. Vapor pressure: partial pressure exerted by water vapor vapor in air on the earth's surface due to its own weight

II. Absolute humidity: mass of water vapor contained in a unit volume of air at any instant

III. Specific humidity: mass of water vapor per unit mass of moist air

IV. Relative humidity: Actual vapor pressure (e_a)/ Saturation vapor pressure(e_s)

(or ratio of the amount of water vapor actually contained per unit volume to the amount of water vapor that it can hold at the same temperature when saturated)

V. Mixing ratio: mass of water vapor per unit mass of perfectly dry air in a humid mixture

Saturation vapor pressure gradient

Gradient/slope of saturation vapor pressure (e_s) curve is found by differentiating e_s with respect to temperature.

$$\Delta = \frac{de_s}{dT} = \frac{4098e_s}{(237.3 + T)^2}$$

Δ= slope of saturation vapor pressure, T = temperature in ^{0}C

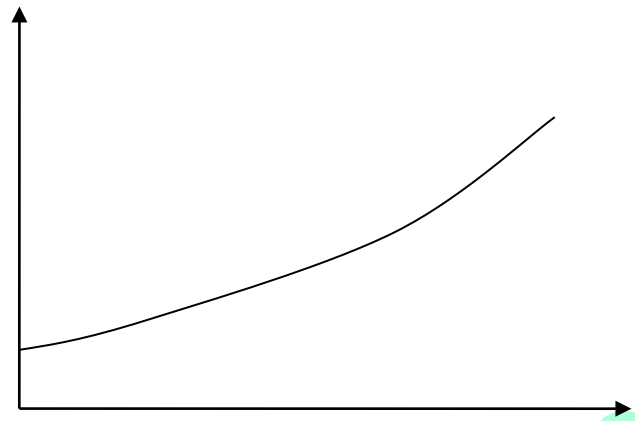


Fig. 3.1: Saturation vapor pressure curve

Dew point temperature
The temperature at which air becomes saturated when cooled at constant pressure and moisture content is called dew point temperature.

Measurement of humidity
a. By using psychrometer: It contains wet bulb thermometer (continuous moisture supply by wrapping with wick and submerging the other end in distilled water) and dry bulb thermometer (recording ambient air temperature).
b. By using hygrograph: Automatic recording of humidity
Principle: hair reacts to the changes in air humidity by expanding or contracting

c. Wind

Wind is a moving air. Wind has both speed and direction. Wind direction is the direction from which it is blowing. Wind speed varies with height above the ground. Wind is one of the major factors that affect the climate and evaporation rate from water surface. Wind influences the ability to transport vapor away from the surface as well as the temperature of the area. Higher wind speed results in higher evaporation rate from a water surface as the wind replaces saturated air just above the water surface by unsaturated air.

Wind speed is measured by anemometers. For comparable data, all anemometers are installed at same elevation above ground. Wind speed varies greatly with height above the ground due to ground friction, trees, buildings and other obstacles.

Wind speed at a certain height is computed by power law as

$$\frac{V}{V_0} = \left(\frac{Z}{Z_0}\right)^{0.15}$$

V = velocity at any height Z
V_0 = Observed velocity at height Z_0

Types of wind

a) Sea and land breezes: See breeze is the blowing of wind from sea to land due to higher temperature (lower atmospheric pressure) at land during day time. Sea breeze is the reason we feel cooler near large water body at day time in a hot day. Land breeze is the blowing of wind from land to sea due to quicker cooling of land, and hence denser air above land surface.
b) Monsoon (seasonal) Winds: Winds whose direction depends on season.
c) Cyclone (hurricane/typhoon): Cyclones are caused when a low pressure area is surrounded by high pressure areas around which air flows anticlockwise in the northern hemisphere and clockwise in southern hemisphere. A cyclone is generally followed by heavy rain.
d) Anticyclone: Anticyclone is a region of high pressure surrounded by low areas around which air flows clockwise in the northern hemisphere and anticlockwise in southern hemisphere.
e) Tornadoes: Tornadoes are similar to cyclone, but they generally form over ocean. Tornadoes are generally destructive to land and property.
f) Local winds: They affect only limited areas and blow for short durations. The cause of local winds is mostly local temperature depressions.

d. Radiation

Radiation is the direct transfer of energy by means of electromagnetic waves. Radiation from the sun is called solar radiation. Solar Radiation provides the fuel for the hydrologic cycle. Solar radiation determines weather and climate of earth.

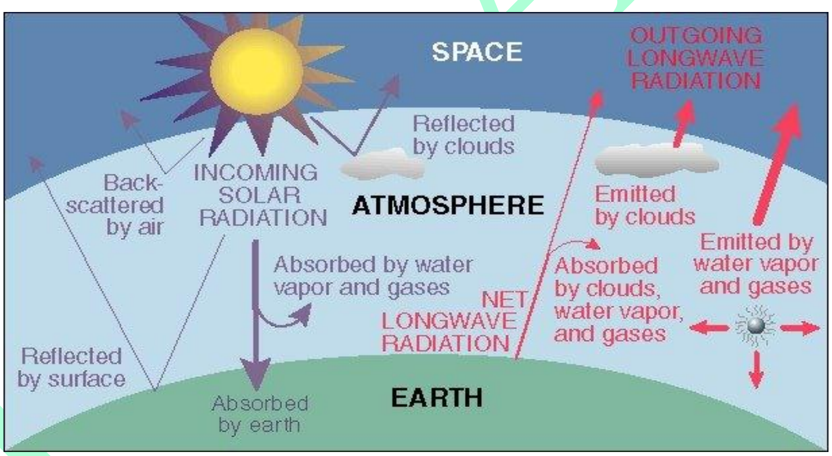


Fig. 3.2: Components of radiation balance

Terminology

Insolation: incident solar radiation

Short wave and long wave radiation
Solar radiation from the sun is referred to as short wave radiation. The radiation from the earth is referred to as long wave radiation.

Albedo
When radiation strikes a surface, it is either reflected or absorbed. The ratio of amount of solar radiation reflected by a body to incoming radiation is called albedo.

Net radiation
The net radiation is the difference between absorbed radiation and emitted radiation.

$$r = \frac{R_r}{R_i}$$

$$R_r = rR_i$$

r = albedo
R_r= Reflected radiation
R_i =Incoming radiation

Absorbed radiation=$R_i - rR_i = R_i(1-r)$
Emitted radiation =R_e
Net radiation = $R_i(1-r) - R_e$

Radiation emission is governed by Stefan-Boltzmann law:

Radiation is also continuously emitted from all bodies at rates depending on the temperatures.

$$R_e = e\sigma T^4$$

E = emissivity of the surface
σ= Stefan-Boltzmann constant
T = Absolute temperature of the surface (K)

Net radiation at the Earth's surface
Incoming radiation = Long wave radiation (R_l) + Shortwave radiation (R_s)
Radiation emitted by the earth = R_e
Albedo = r
Net radiation at the Earth's surface (R_n) is given by

$$R_n = (R_l + R_s)(1-r) - R_e$$

Intensity of solar radiation depends on

- Scattering in the atmosphere
- Absorption by clouds
- Obliqueness of the Earth's surface to the incoming radiation (a function of latitude, season and time of day)

Radiation measurement

Actinometers and radiometers are used to measure intensity of radiant energy. The data is used in studies of evaporation and snowmelt.

3.2.2 Factors affecting evaporation

I. Meteorological factors

- Radiation: most important factor as it directly influences the temperature of the evaporating surface.
- Temperature: Increase in temperature increases the evaporation rate but not always proportionally. For same temperature, evaporation in colder months is less than summer months due to other environmental factors.
- Humidity: Humidity influences vapor pressure deficit which governs the rate of evaporation.
- Vapor pressure: Evaporation is proportional to the difference between saturation vapor pressure at the water temperature and actual vapor pressure in the air.
- Wind: Wind helps to carry away moisture as it evaporates and thus accelerates the rate of evaporation. Generally the rate of evaporation increases with the wind speed up to a critical speed beyond which any further increase in wind speed has no influence on the rate of evaporation. There is a relation between wind speed and size of water surface.
- Atmosphere pressure: Increase in atmosphere pressure decreases the rate of evaporation.

II. Nature of evaporating surface

- Soil: the rate of evaporation from soil depends on the availability of water, e.g, higher rate for wet soil, lower rate for dry soil.
- Snow and ice: Evaporation from snow can occur when the vapor pressure of the air is less than that of the snow surface i.e. only when the dew point is lower than the temperature of the snow.
- Reservoir: The rate of evaporation from a reservoir depends on the heat storage capacity, e.g. for deep water bodies, large heat storage during summer causing less evaporation and vice versa in winter.

III. Quality of water: Soluble salts reduce the vapor pressure, and thus reduce the rate of evaporation.

3.2.3 Methods of estimation of evaporation

a. Empirical equations

Empirical equations used for estimating evaporation are functions of saturation vapor pressure at the water temperature (e_s) and actual vapor pressure in the air (e_a).

General equation (Dalton's law): $E = kf(u)(e_s - e_a)$
E= evaporation
k = coefficient
f(u) = wind speed correction function
e_s = saturation vapor pressure
e_a = actual vapor pressure

Meyer's formula

$$E = C\left(1 + \frac{U}{16}\right)(e_s - e_a)$$

E= Evaporation (mm/day)
U= monthly mean wind speed in km/h measured at 9m above ground
C = coefficient (0.36 for large lakes, 0.50 for shallow lakes)
e_s = saturation vapor pressure (mm of Hg)
e_a = actual vapor pressure (mm of Hg)

Rhower's formula
$E = 0.771\ (1.465-0.000732P)\ (0.44+0.0733U)\ (e_s-e_a)$
E= Evaporation (mm/day)
P= mean barometric reading in mmHg
U = mean wind velocity at 0.6m above ground in km/h
e_s = saturation vapor pressure (mm of Hg)
e_a = actual vapor pressure (mm of Hg)

b. Analytical methods

I. Water budget method

∑Inflow -∑Outflow =Change in storage + Evaporation loss

$$E = \sum I - \sum O \pm \Delta S$$

General equations
$E= (P+I_{sf}+I_{gf})-(O_{sf}+O_{gf}+T)\pm\Delta S$

P= precipitation
I_{sf} = Surface inflow
I_{gf} = Groundwater inflow
O_{sf} = Surface water outflow
O_{gf} = Groundwater outflow
T = Transpiration loss
ΔS = Change in storage

- Measurement of I_{gf}, O_{gf} and T is not possible, these can only be estimated.
- T is usually negligible.
- Water budget equation gives approximate values.

II. Energy budget method

Based on law of conservation of energy
Incoming energy = outgoing energy + Change in stored energy

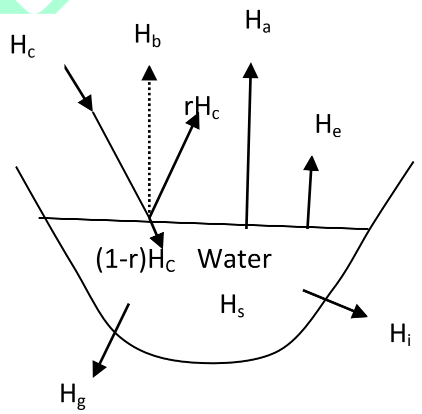


Fig. 3.3: Components of energy balance

Energy balance to evaporating surface in a period of one day
$H_n = H_a + H_e + H_g + H_s + H_i$
H_n = net radiation = $H_c(1\text{-}r)\text{-}H_b$
r H_c = Reflected radiation
H_c= Incoming solar radiation
R =albedo
H_b = Back (Long wave) radiation from water body
H_a = Sensible heat transfer from water surface to air
H_e = Heat energy used up in evaporation
H_g = Heat flux into the ground
H_s = Heat stored in water body
H_i = Net heat conducted out of the system by water flow (advected energy)

For short time period H_s and H_i can be neglected. All the terms except H_i can either be measured or evaluated indirectly. H_a is estimated using Bowen's ratio.

The ratio of sensible heat flux to heat flux used up in evaporation is called Bowen ratio.

$$\beta = \frac{H_a}{H_e} = \frac{H_a}{\rho . E . L}$$

β = Bowen ratio
ρ = Density of water
E = Evaporation
L = Latent heat of vaporization
Estimate of β

$$\beta = \gamma \frac{Tw - Ta}{ew - ea}$$

γ = Pshcrometric constant
ew = saturated vapor pressure (mmHg)
ea = actual vapor pressure (mmHg)
Tw = Temperature of water surface(C)
Ta = Temperature of air (C)

$$E = \frac{H_n - H_g - H_s - H_i}{\rho L(1+\beta)}$$

III. Mass transfer method

This method is based on theories of turbulent mass transfer in boundary layer to calculate the mass water vapor transfer from the surface to the surrounding atmosphere.

3.2.4 Evaporimeters (Evaporation pan)

Evaporation Pan, also called Evaporimeter, is shallow vessels containing water. These are placed in open to measure the loss of water by evaporation. Water is placed in the evaporation pan and the change in depth of water due to evaporation is measured.
Lake or reservoir evaporation = Pan coefficient x Pan evaporation

Pan Evaporation differs from lake evaporation due to the depth of exposure of pan above ground, color of the pan, height of the rim, heat storage and heat transfer capacity with respect to reservoir, variation in vapor pressure, wind speed and water temperature. Pan coefficient takes into account these factors.
Pan coefficient: 0.6 to 0.8

Various types of pans
Class A evaporation pan
It consists of a cylindrical vessel made of galvanized iron sheet. The pan is placed 15cm above the ground surface in such a way that it gets free circulation of air.

Sunken pan (Colorado Sunken pan)
The pan is buried into the ground such that the water level is at the ground level. Advantage of this pan is that the aerodynamic and radiation characteristics are closer to the reservoir. The water level is maintained at or slightly below the ground level.

3.3 Evapotranspiration

The processes of evaporation from the land surface and the transpiration from the vegetation are collectively termed evapotranspiration (ET).
Main factors affecting ET

- Supply of energy (solar radiation)
- Ability to transport vapor away (wind speed and humidity gradient)
- Supply of moisture at the evaporating surface

Potential Evapotranspiration and Actual Evapotranspiration

Potential Evapotranspiration (PET) is the evapotranspiration that would occur from a well vegetated surface when moisture supply is not limiting. The real evapotranspiration occurring in a specific situation is called actual evapotranspiration (AET).

Field capacity and permanent wilting point

Field capacity is the maximum quantity of water that the soil can retain against the force of gravity. Permanent wilting point is the moisture content of a soil at which the moisture is no longer available in sufficient quantity to sustain the plants. The difference in these two moisture contents is called available water.

If the water supply to the plant is adequate, soil moisture will be at filed capacity and AET = PET.
If the water supply is less than PET, the soil dries out and AET<PET.
At permanent wilting point, AET = 0

Penman method for determination of evapotranspiration

Penman method is a combined aerodynamic and energy balance method for estimating evapotranspiration. Evapotranspiration is computed by aerodynamic method when energy supply is not limited and by the energy balance method when vapor transport is not limited. But, normally, both of these factors are limiting, so a combination of the two methods is needed.

Assumptions:

- Steady state energy flow prevails.
- Changes in heat storage over time in the water body are not significant.
- Vapor transport coefficient is a function of wind speed.
- Advected energy input is small, which may be neglected.

Penman's formula for estimation of evapotranspiration is given by

$$PET = \frac{AH_n + \gamma E_a}{A + \gamma}$$

PET = daily potential evapotranspiration (mm/day)
A = slope of saturation vapor pressure (mmHg/^{0}C)
H_n =Net radiation (mm/day)
E_a = Evaporation due to aerodynamic method (mm/day)
γ = Psychrometric constant (mmHg/^{0}C) (can be taken as 0.49mmHg/^{0}C)

The net radiation is estimated by the following equation:

$$H_n = H_a(1-r)\left(a + b\frac{n}{N}\right) - \sigma T_a^4\left(0.56 - 0.092\sqrt{e_a}\right)\left(0.10 + 0.90\frac{n}{N}\right)$$

H_a = Incident solar radiation outside the atmosphere on a horizontal surface (mm/day)
a = constant depending upon latitude φ and is given by a = 0.29Cos φ
b = constant with an average value of 0.52
n = actual duration of bright sunshine hours (hours)
N = Maximum possible hours of bright sunshine (hours) (function of latitude)
r = albedo
σ = Stefan-Boltzman constant = 2.01×10^{-9} mm/day
T_a = mean air temperature (degree Kelvin) = 273+^{0}C
e_a = Actual vapor pressure (mmHg)

E_a is estimated as

$$E_a = 0.35\left(1 + \frac{u2}{160}\right)(e_s - e_a)$$

u2 = mean wind speed at 2m above ground (km/day)
e_s = Saturated vapor pressure at mean air temperature (mmHg)
e_a = Actual vapor pressure (mmHg)
For the computation of PET, data on temperature, wind speed, radiation (or sunshine hours) and vapor pressure (or humidity) are needed. H_a, N and A are obtained from tabulated values, or from equations.

Value of e_s from T
$e_s = 4.584\, exp\left(\frac{17.27T}{237.3+T}\right)$
e_s = saturation vapor pressure (mmHg)
T= Temperature (^{0}C)
If Relative humidity (RH) is given, RH = e_a/e_s
Equation to compute A
$A = \frac{4098 e_s}{(237.3+T)^2}$ where e_s = saturation vapor pressure (mmHg), T= Temperature (^{0}C)
For 20^{0}C, A = 1.08 mmHg/^{0}C

Value of r: Water surface = 0.05, Bare land: 0.05-0.45

Measurement of evapotranspiration

Lysimeter Method

Lysimeter is a small tank containing soil in which the plants are grown. It is generally cylindrical tank about 60 to 90 cm in diameter and 180 cm deep. This tank is buried in ground such that its top is made like the surrounding ground surface. Water is applied to the lysimeter for the satisfactory growth of plant. Percolated water excess to the plant use is collected in a pit and Evapotranspiration is obtained.

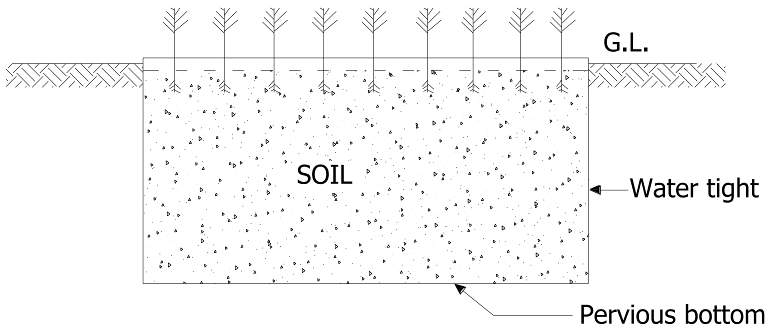


Fig. 3.4: Lysimeter

Computation of evapotranspiration

$$P+W=O+ET+\Delta S$$

P = precipitation

W = Amount of water applied

O = Quantity of water drained out

ET = Evapotranspiration

ΔS = Change in soil moisture storage

3.4 Infiltration

3.4.1 Introduction

Infiltration is the process by which water enters the soil from the ground surface. Infiltration first replenishes the soil moisture deficiency. The excess water then moves downwards by the force of gravity. This downward movement under gravity is called percolation (or seepage). Percolation is thus the movement of water within the soil.

Infiltration rate (f) is the rate at which water enters the soil at the surface. Cumulative infiltration (F) is the accumulated depth of water infiltrated during a given time period.

$$F(t) = \int_{o}^{t} f(t)dt$$

$$f(t) = \frac{dF(t)}{dt}$$

Infiltration capacity (fc) is the maximum rate at which a given soil can absorb water under a given set of conditions at a given time.

The actual rate of infiltration (f) can be expressed as

$f = fc$ for $i \geq fc$

$f = i$ for $i < fc$

i = intensity of rainfall

Infiltration capacity of a soil is high at the beginning of a storm and has an exponential decay as the time elapses.

Hydraulic conductivity: It is a measure of ability of the soil to transmit water.

Field capacity: Field capacity is the maximum amount of water that the soil can hold against the force of gravity.

Moisture zones

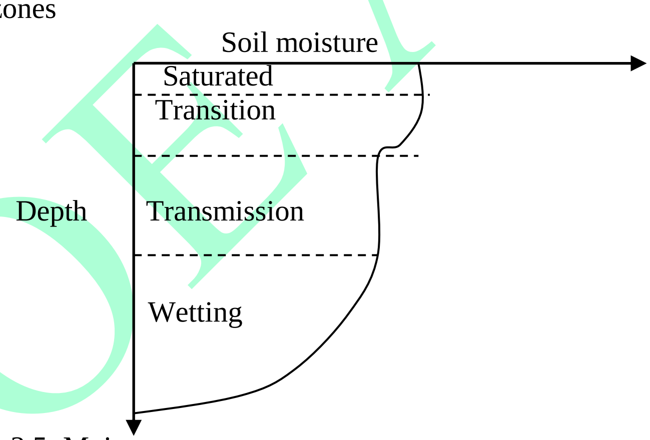


Fig.3.5: Moisture zones

- Saturated zone: top zone
- Transition zone: second zone
- Transmission zone: uniform moisture content, moisture content above field capacity but below saturation, unsaturated zone
- Wetting zone: moisture content at or near field capacity, decrease of moisture with depth, wetting front as sharp discontinuity

Factors affecting infiltration (f)

1. Characteristics of soil
   - Type of soil, Porosity, texture (determines size of pores), Structure (affects aggregation)
   - Permeability: high f for loose, permeable sandy soil
   - Underdrainage: high f for good underdrainage
   - Grain size of soil particles: higher f for large grain size
   - layering
2. Condition of soil surface and its vegetative cover
   - Low f for bare soil: Clogging the surface by inwashing of fine particles
   - Grass and vegetation cover: high f
3. Antecedent moisture content of the soil
   - Second storm in succession: low f
4. Climatic conditions
   - Temperature affecting viscosity and thus f (less viscous, more f)
5. Rainfall intensity and duration
   - Intense rainfall: progressive reduction of f due to increased supply of moisture, mechanical compaction and in-wash of finer particles
   - Sustained heavy rainfall of longer duration: steady reduction in f_c until f attains a constant value.
6. Human activities
   - Crop growing: increase of f
   - Construction of road, house etc.: reduction of f
7. Quality of water
   - Presence of salt: affecting viscosity and reducing porosity, lower f
   - Turbidity: clogging pore space, lower f
8. Groundwater table
   - Close to surface: low f

3.4.2 Horton equation for infiltration

According to Horton, Infiltration begins at some rate f_0 and exponentially decreases until it reaches a constant value f_c

$$f(t) = f_c + (f_0 - f_c)e^{-kt}$$

f(t): infiltration capacity at any time t from the start of the rainfall
f_0: initial infiltration capacity at t = 0
f_c : infiltration rate at the final steady stage when the soil profile becomes fully saturated
k : decay constant depending upon soil characteristics and vegetation cover, known as Horton coefficient
Three parameter to fix: f_0, f_c, k, practical difficulty in determination

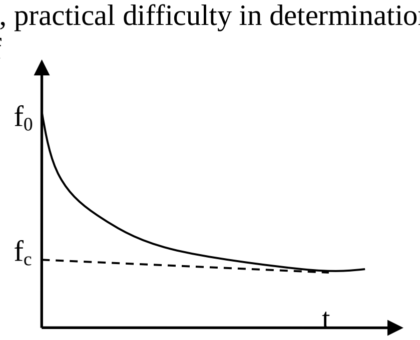


Fig. 3.6 : Infiltration curve

Cumulative infiltration or total infiltration using Horton's equation for time t from start

$$F(t) = \int_{o}^{t} f(t)dt$$

$$= \int_{o}^{t} [f_c + (f_0 - f_c)e^{-kt}]dt$$

$$= f_c t + (f_0 - f_c)\left|\frac{e^{-kt}}{-k}\right|_0^t$$

$$F(t) = f_c t + \frac{f_0 - f_c}{k}\left(1 - e^{-kt}\right)$$

Average infiltration in time t = F(t)/t = $f_c + \frac{f_0 - f_c}{kt}\left(1 - e^{-kt}\right)$

Cumulative infiltration or total infiltration depth in between time t1 and t2

$$F(t) = \int_{t1}^{t2} f(t)dt$$

$$= \int_{t1}^{t2} [f_c + (f_0 - f_c)e^{-kt}]dt$$

$$= f_c(t2 - t1) + \frac{f_0 - f_c}{(-k)}(e^{-kt2} - e^{-kt1})$$

To determine k with known values of F(t), f_c, f_o and t

For large t, the value of e^{-kt} becomes negligible. Hence above equation reduces to

$$F(t) = f_c t + \frac{f_0 - f_c}{k}$$

$$k = \frac{f_0 - f_c}{F(t) - f_c t}$$

If rainfall intensity (i) is less than f, all rainfall is infiltrated. Runoff occurs only after i>f.

Determination of constants f_0, f_c and K from given data of f and t

a. Graphical approach

Plot f on Y-axis and t on x-axis. Draw exponential curve and note down the values of f_0 and f_c.
Horton's infiltration equation is given by
$f = f_c + (f_0 - f_c)\, e^{-Kt}$
$f - f_c = (f_0 - f_c)\, e^{-Kt}$
Integrating

$$\int_0^{\infty} (f - f_c)dt = F = Area\ under\ the\ curve$$

$$\int_0^\infty (f_0 - f_c)\, e^{-kt} dt = \frac{f_0 - f_c}{k}$$

Equating above expressions, K can be determined by

$$k = \frac{f_0 - f_c}{F}$$

b. Statistical approach

Horton's infiltration equation is given by
$f = f_c + (f_0 - f_c)\, e^{-Kt}$
$f - f_c = (f_0 - f_c)\, e^{-Kt}$
Taking log on both sides
$\ln(f - f_c) = \ln(f_0 - f_c) - Kt$
Let $y = \ln(f - f_c)$, $c = \ln(f_0 - f_c)$. Then above equation reduces to
$y = -Kt + c$: linear equation

Procedure:
Take f_c from the given data.
Determine K and C by least square method.
$K = -\frac{N\sum xy - \sum x \sum y}{N\sum x^2 - (\sum x)^2}$
$c = \frac{\sum y - (-K)\sum x}{N}$
With $c = \ln(f_0 - f_c)$, compute f_0.

Excess rainfall or effective rainfall

3.4.3 Infiltration indices

Average rate of infiltration is called infiltration index. For computation of surface runoff and flood discharge, the use of infiltration curve is not convenient. So, we can use constant value of infiltration rate for the duration of storms.

Two common infiltration indices
a. φ index

The average rate of rainfall above which the rainfall volume equals to runoff volume is called φ index. It is based on the assumption that for a specified storm with given initial conditions, the rate of basin recharge remains constant throughout the storm period. i.e. φ remains constant.
For $i < \varphi$, $f = i$
For $i > \varphi$, runoff = $i - f$
i = rainfall intensity
f = infiltration rate
φ : total abstractions
The amount of rainfall in excess of the index is known as effective rainfall or rainfall excess.

Method of determination of φ index
Given: rainfall hyetograph and direct runoff
Use same unit ,e.g. mm, cm for rainfall and runoff.
Take incremental rainfall if cumulative rainfall is given.

Method 1

Trial and error with effective time (t_e)

a. Consider the whole duration of rainfall as effective in the beginning.

First trial: φ= (Total rainfall - Direct runoff)/t_e

where t_e = total time of excess rainfall contributing for direct runoff (effective duration)

b. Compute rainfall excess of each rainfall pulse and find total rainfall excess.

Rainfall excess = observed rainfall (R) - φ Δt where Δt = interval of rainfall data for rainfall intensity> φ, 0 otherwise

c. Compare total rainfall excess with direct runoff. If rainfall excess (Re) is not same as direct runoff (Q), take another value of te. Take te by subtracting ineffective rainfall duration from whole period.

Second trial: φ= (Total rainfall - Direct runoff-Ineffective rainfall)/t_e

d. Repeat steps b-c until Re=Q.

Method 2

Trial and error with φ

a. Consider the whole duration of rainfall as effective in the beginning.

First trial: φ= (Total rainfall - Direct runoff)/t_e

where t_e = total time of excess rainfall contributing for direct runoff (effective duration)

b. Compute rainfall excess of each rainfall pulse and find total rainfall excess.

Rainfall excess = observed rainfall (R) - φ Δt where Δt = interval of rainfall data for rainfall intensity> φ, 0 otherwise

c. Compare total rainfall excess with direct runoff. If rainfall excess (Re) is not same as direct runoff (Q), take another value of φ.

Re>Q, increase φ

Re<Q, decrease φ

d. Repeat steps b-c until Re=Q.

2. W-index

A w-index is defined as average rate of infiltration during the time rainfall intensity exceeds the infiltration capacity. This index is considered as an improvement over φ index in the sense that initial losses (interception and surface storage) are considered.

$$W = \frac{P - R - I_a}{t_e}$$

P = total storm precipitation

R = Total storm runoff

I_a = Initial losses

t_e = duration of the excess rainfall (time during which rainfall rate exceeds infiltration rate)

W =average rate of infiltration

For I_a = 0 (long and heavy storms) φ index = W index

Determination of I_a is difficult. So W_{min} index is used instead of W-index when the soil condition is very wet so that the soil infiltration rate is almost constant and infiltration is at the minimum rate for the basin.

φ index and W index depends on soil type, vegetal cover, initial moisture condition, and storm duration and intensity.

Determination of W index

Prepare the rainfall data by deducting the initial loss from the first pulse of rainfall and then follow the same procedure as that of φ index.

3.4.4 Measurement of infiltration

1. Ring infiltrometer

Ring infiltrometer is a metal ring that is driven into the soil. There are two types of infiltrometers:

a. Single tube infiltrometer: It is a hollow metal cylinder of 60cm long and 30 cm in diameter. Water is placed inside the ring and the level of water is recorded at regular time intervals as it recedes. This data is used to prepare cumulative infiltration curve, from which infiltration capacity as a function of time may be calculated

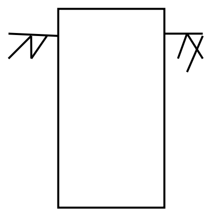


Fig. 3.7: Simple infiltrometer

b. Double tube infiltrometer: It consists of two concentric hollow cylinders of same length. Water is added to both rings to maintain the same height. The infiltration data from the inner cylinder is taken as infiltration capacity of the soil. The outer cylinder is maintained to prevent spreading of water from the inner one.

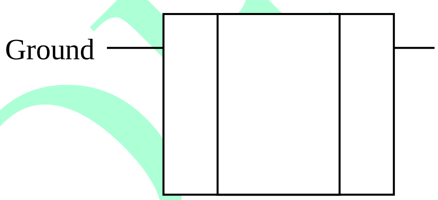


Fig.3.8: Double tube infiltrometer

2. Rainfall simulator

A rainfall simulator consists of a sprinkler with nozzles capable of producing artificial rain of various intensities, drop sizes and durations. A field plot of about 2mx4m is selected on which the nozzles spray water at a height of 2 m or more to the field. Arrangement is made to collect and measure the runoff from the plot. Experiments are conducted under controlled conditions with various combinations of intensities and durations. Using the water budget equation, infiltration rate is estimated.

$F_d = P_d - S_{rd} - S_{ol}$

F_d = Depth of infiltrated water

P_d = Simulated rainfall depth
S_{rd} = Surface runoff depth
S_{ol} = other losses, e.g. depression storage, detention, abstraction

Chapter 4:Surface runoff

Characteristics of drainage basin

Drainage basin/watershed/catchment
Basin area (A): area of land draining into a particular location of a stream

For delineating basin, we need topo map. The map shows changes in elevation by using contour lines.
Features of contour

- Uphill: contour with higher elevation
- Hill: circular contour, ridge: highest point
- Saddle: mountain pass
- Valley: V or U shaped with the point of the V/U being the upstream end
- Close together contours: steep slope
- Widely spaced contour: level ground

Basin delineation procedure on topo map

- Mark the outlet point
- Mark the highest point (ridge line: catchment divider) around the river
- Start from the outlet and draw line perpendicular to the contours in such a way that the line passes from the highest point (ridge)
- Continue to the opposite side of the watercourse, finally ending to the outlet.

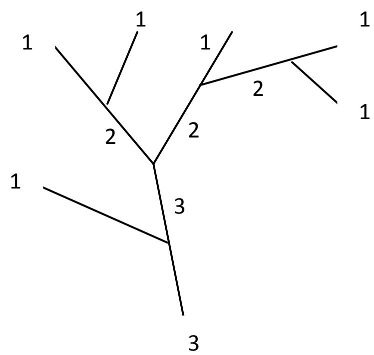
Relation of watershed discharge Q with basin area A: $Q = xA^y$

Stream order

- measure of amount of branching within a stream

Stream order assigning procedure

- The smallest recognizable channels are designated order 1. (non-branching tributary)
- Where two channels of order 1 join, a channel of order 2 results downstream (receiving flow from 1^{st} order). In general, where two channels of order i join, a channel of order i+1 results.
- Where a channel of lower order joins a channel of higher order, the channel downstream retains higher of the two orders.
- Order of the basin: order of the stream draining at outlet = highest order in the basin



Example of stream order

Variables based on stream ordering

Bifurcation ratio (R_B): ratio of the number N_i, of channels of order i to the number N_{i+1} of channels of order i+1

$R_B = N_i/N_{i+1}$

R_B: relatively constant from one order to another

Length ratio (R_L): ratio of average length of streams of order i+1 to that of order i

$R_L = L_{i+1}/L_i$

Area ratio (R_A): ratio of average area drained by streams of order i+1 to that of order i

$R_A = A_{i+1}/A_i$

Drainage density (D_d): ratio of total length of all streams of the basin to its area

$D_d = L_s/A$

- Indication of drainage efficiency
- Higher D_d, quicker runoff, less infiltration and other losses

Length of overland flow = $1/(2\ D_d)$

Length area relationship (Horton's formula): $L = 1.4\ A^{0.6}$ where A- mile^2 (Useful for large rivers of the world), L -mile

Stream density (D_s): ratio of number of streams of given order per sq. km.

$D_s = N_s/A$

Shape of the basin

Shape of the basin governs the rate at which water enters the stream. The shape of basin is expressed by form factor.

Form factor = average width of basin (B)/axial length of basin (L) = A/L^2

Shape factor (B_s): ratio of square of basin length (L) to its area (A), $B_s = L^2/A$

Slope of the Channel

Slope of channel affects the velocity and flow carrying capacity at any given location at its course.

Slope = elevation difference between 2 points of a channel(h)/horizontal length between points (L)

Centroid of basin

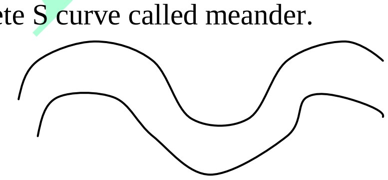
Location of point of weighted center

Hydraulic geometry

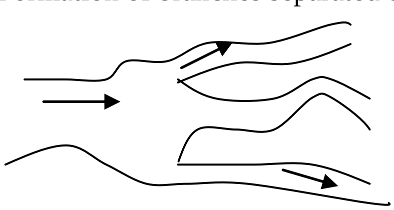
It includes the character of channel, longitudinal variation of mean depth, width and velocity at a particular cross-section.

Stream pattern

a) **Meandering types -** Formation of successive bends of reverse order leading to the formation of a complete S curve called meander.



b) **Braided -** Formation of branches separated by islands



c) **Straight** - Straight and single channel.

Flood plains
The flood plains of a river are the valley floor adjacent to the channel, which may be inundated during high stage of river. Flood plains are formed due to the deposition of sediment in the river channel and deposition of fine sediments on the flood plains on flooding.

4.10 Factors affecting runoff

Physiographic factors	Climatic factors
1. Basin characteristics a. Shape b. Size c. Slope d. Nature of valley e. Elevation f. Drainage density 2. Infiltration characteristics a. land use and cover b. soil type and geological conditions c. lakes, swamps and other storage 3. Channel characteristics: cross-section, roughness and storage capacity	1. Strom characteristics Precipitation: duration, intensity and magnitude Movement of storm 2. Initial loss 3. Evappotranspiration

Shape
- Time taken for the water to reach to outlet from remote part depends upon the shape of basin.
- Fan shaped: Greater runoff (same size tributaries, almost similar time of concentration)
- Elongated: broad and low peak (distributed over time)
- Peak flow proportional to square root of drainage area

Size
- Small basin: overland flow predominant
- Large basins: channel flow predominant, constant minimum flow than small basins

Slope
- Slope: control velocity of flow
- Related to overland flow, infiltration capacity and time of concentration of rainfall in streams
- Large stream slope: quicker depletion of storage
- Steeper slope for small basin: higher peak

Elevation
- Affects mean runoff (effect of evaporation and precipitation and effect of snow)

Drainage density
- Drainage density = total channel length/total drainage area
- High density: fast response

- Low density: slow response

Land use

- Vegetal cover: reduce peak flow
- Barren land: high runoff

Soil

- Type of soil and subsoil and their permeability conditions
- Geology: Controls infiltration

Storage: reduce runoff
Lakes: reduce flood

Rainfall intensity: increase in runoff with increase in intensity
Rainfall duration: controls volume of runoff
Rainfall distribution: maximum runoff occurs when entire catchment contributes to runoff.
Direction of the storm movement: affects peak flow and time of duration of runoff, up to down: high runoff, down to up: low runoff
Evapotranspiration: inversely proportional to runoff

Rainfall runoff relationship

Correlation: degree of association between variables.

The relationship between rainfall and runoff is complex due to a number of factors. Therefore, simple method like correlating runoff with rainfall is used in practice.

a. Linear
Equation for straight line regression
R = aP + b
where R = Runoff
P = Rainfall
a, b: constants

Coefficients by regression

$$a = \frac{N(\sum PR) - (\sum P)(\sum R)}{N(\sum P^2) - (\sum P)^2}$$

$$b = \frac{\sum R - a \sum P}{N}$$

N = number of observations
Coefficient of correlation

$$R = \frac{N(\sum PR) - (\sum P)(\sum R)}{\sqrt{[N(\sum P^2) - (\sum P)^2][N(\sum R^2) - (\sum R)^2]}}$$

b. Exponential
For large catchments, exponential relationship can be developed

$$R = \beta P^m$$

β, m: coefficients
Taking log for linearization

$$lnR = mlnP + ln\beta$$

With R = lnR, a = m, P = lnP, b = ln β, above equation reduces to the one same as before.

4.1 Stream gauging

Streamflow

- That part of precipitation which appears in a stream as surface runoff.

Discharge

- Volume of water flowing through a channel cross section per unit time.

Hydrometry: science of measurement of water

Stage or gauge height

- The elevation of water surface at a location in any water body above a reference datum.

Water body: River, lake, canal, reservoir

Stream gauging station

Stream gauging station is defined as the location at which the river discharges are recorded and the discharge measurements are carried out.

Purpose of stream gauging: to provide systematic records of stage and discharge

Factors to be considered for the selection of site for stream gauging

- Easily accessible
- Stable and fairly straight river reach about 100m u/s and d/s
- Stable and regular channel bed
- Free from backwater effects
- Regular cross section
- No excessive turbulence and eddies
- No excessive vegetal or aquatic growth
- Velocities: neither too high nor too low, generally in the order of 0.1-5m/s

4.2 Stage measurement

1. Manual or non-recording gauge

Manual gauge is read and recorded by observer/gauge reader once, twice, thrice daily or more. It does not provide continuous record of stage. It is cheaper and easier to install.

Staff gauge

Staff gauge is the most common and simplest form of manual gauge. It consists of a graduated plate fixed in the stream or on the bank of river or on a structure e.g. bridge abutment or pier. The level of water surface in contact with the gauge is measured by matching the reading of the staff and adding with reference datum level.

It is of three types

- Vertical: one vertical gauge
- Sectional: more than one gauges at different locations
- Inclined

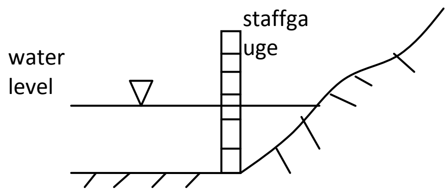


Fig. 4.1: Staff gauge

2. Recording gauge or automatic gauge

Recording gauge records continuous stage of a river over time.

Two common automatic gauges
a. Float gauge

A float is connected to one end of a wire which passes through a recorder, and the other end of a rope is balanced by a suitable counterweight. Displacement of float due to rising or lowering of water level causes an angular displacement of pulley and hence of the input shaft of the recorder. Mechanical linkages convert this angular displacement to the linear displacement of a pen to record over a drum driven by clockwork. The float gauge is protected by installing a stilling well.

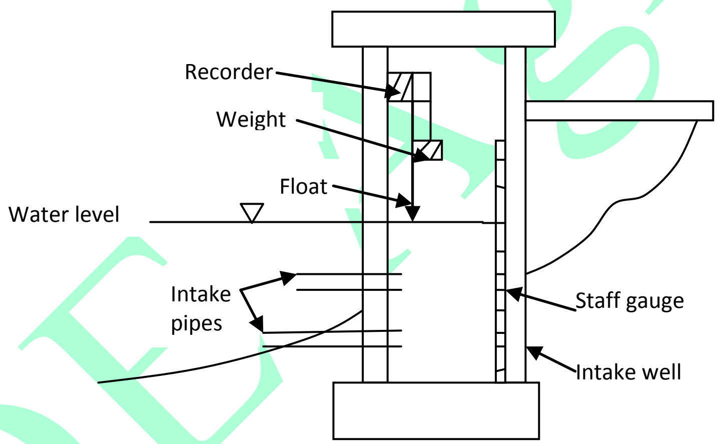


Fig. 4.2: Arrangement of float gauge

b. Bubble gauge

Bubble gauge consists of small tube placed at the lowest water level through which compressed air (usually CO_2 or N_2 gas) is continuously bubbled out. The pressure required to continuously push the gas stream out beneath the water surface is a measure of depth of water over the nozzle of the bubble stream. This pressure is measured by a manometer in the recorder house.

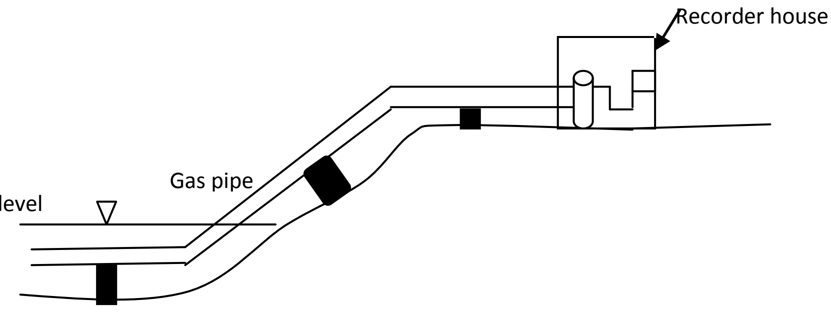


Fig. 4.3: Arrangement of bubble gauge recorder

4.3 Discharge measurement using velocity-area method

Velocity area method

This involves the measurement of velocity at the gauging site and the corresponding discharge to obtain river discharge. The velocity is zero at the periphery and changes rapidly as we move from the bank. So a single area-velocity measurement for the entire cross-section will give highly erroneous results. Therefore, the cross-section of a river is divided into a number of subsections by imaginary verticals.

Criteria for selection of verticals

- Each vertical should not pass more than 10% of total discharge
- Width of each subsection = 5% of total width of river
- Difference of velocities in adjacent segments: not more than 20%
- Discharge variation between adjacent subsections: between 5% to 10%

For computation of area, the depth of flow is determined by following methods:

- Wading or sounding rod: If the river be crossed, a wading rod is used to measure the depth of flow. A man walks across the river section with a graduated wading rod to measure water depth.
- Cableway: For deep rivers, cableway is constructed to measure depth and velocity. The lower end of a cable attached to a current meter with a sounding weight is lowered from cablecar. By measuring the length of cable lowered, the depth of flow is measured while velocity is recorded simultaneously by current meter.
- Echo-sounder: In this method, high frequency sound wave is sent down by transducer kept immersed at the water surface and the echo reflected by the bed is also picked up by the same transducer. By comparing the time interval between the transmission of the signal and the receipt of its echo, the distance to the bed is obtained. This method is useful for high velocity streams, deep streams and mobile or soft bed streams.

Velocity is measured by current meter or floats.

Measurement procedure

- Divide the cross-section of the river into n number of verticals.
- At each vertical, measure the horizontal distance from the reference bank, the depth of water and the velocity at one or more points.
- Compute width, cross-sectional area and average velocity to get discharge at each sub-section.

Computation of average velocity in a vertical

- One point method: for depth<1.0 m
  $V_{av} = V_{0.6d}$
- Two point method:
  $V_{av} = 0.5(V_{0.2d}+V_{0.8d})$
  d = depth from water surface

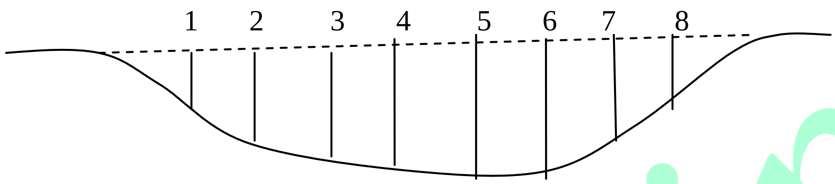


Fig. 4.4: section for area-velocity method

Computation of discharge

I. Mid section method: widely used
In this method, half width to the left and half width to the right of a vertical is taken as width for a sub-section.
For section 2 to n-1
Width (Wav_i) = $\frac{1}{2}(W_i + W_{i+1})$
W_i = Width of section i and W_{i+1} = width of section i+1
For first and last triangular sections

$$Wav_1 = \frac{\left(W_1 + \frac{W_2}{2}\right)^2}{2W_1}$$

$$Wav_n = \frac{\left(W_n + \frac{W_{n-1}}{2}\right)^2}{2W_n}$$

(Alternative way: Width (Wav_i) = $\frac{1}{2}(W_i + W_{i+1})$ can be used from section 1 to n.)

Cross-section area (Ai) = $wav_i.d_i$
whered_i = Depth of section i

Discharge at each subsection (Q_i) = $A_i Vav_i$
Total discharge = $\sum Q_i$

II. Mean section method
Cross-section area

$$A_i = \left(\frac{d_i + d_{i+1}}{2}\right) b_i$$

Discharge at each subsection

$$Q_i = A_i\left(\frac{Vav_i + Vav_{i+1}}{2}\right)$$

Total discharge = $\sum Q_i$

Moving boat method for discharge measurement in a deep river

V_b= Velocity of boat at right angle to the stream
V_f = Flow velocity
V_r = Resultant velocity
θ = angle made by resultant velocity with the direction of boat
Δt = time of transit between two verticals
Convert surface velocity to average velocity. (V_{av} = 0.85xVsurface)

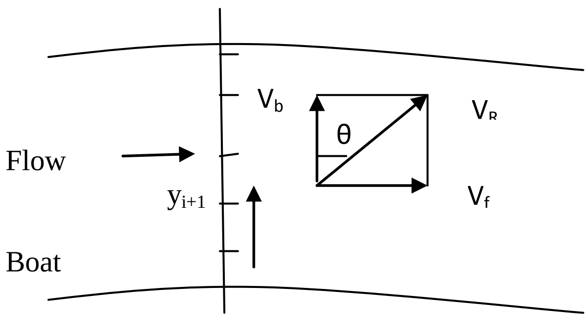


Fig. 4.5: Moving boat method

$V_b = V_R Cos\theta$ and $V_f = V_R Sin\theta$
Discharge in each segment

$$\Delta Q_i = \left(\frac{y_i + y_{i+1}}{2}\right) W_i V_f$$

Where $w_i = V_b \Delta t$
Total discharge = $\sum \Delta Q_i$

4.4 Velocity measurement by current meter

1. Current meter

Current meter is the most commonly used instruments for measuring stream velocity. It consists of a rotating element which rotates due to the reaction of stream current with an angular velocity proportional to the stream velocity. It is weighted down by lead weight called sounding weight to keep in stable position in flowing water.

Types of current meter

a. Vertical axis meter

It consists of a series of conical cups mounted around a vertical axis. The cups rotate in horizontal plane. The revolutions of cup assembly for a certain time is recorded and converted to stream velocity. The normal range of velocity measured by such current meter is 0.15m/s to 4m/s. This type of current meter cannot be used if the vertical component of the velocity is signigicant.

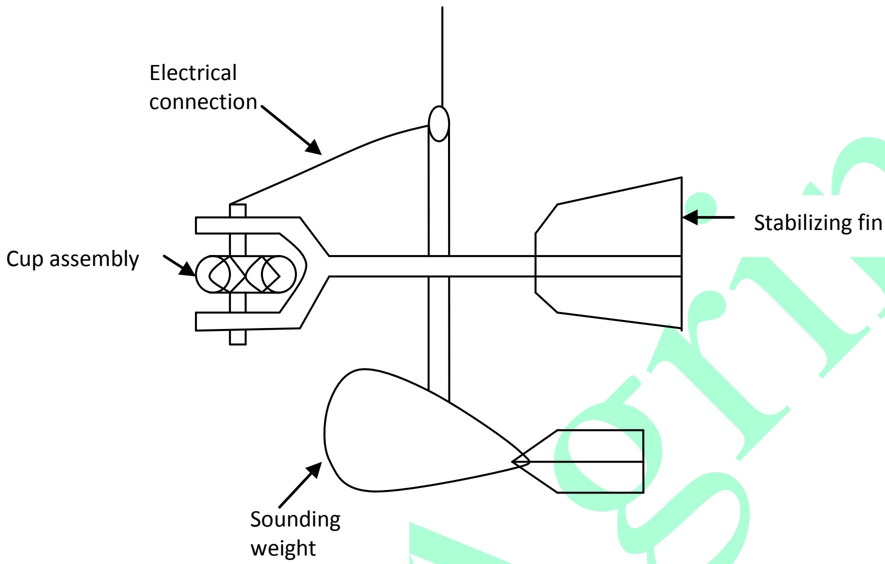


Fig. 4.6: Vertical axis current meter

b. Horizontal axis meter

It consists of a propeller mounted at the end of horizontal shaft.The revolutions of propeller for a certain time is recorded and converted to stream velocity. The current meter can measure velocity from 0.15m/s to 4m/s. This type of current meter is fairly rugged and is not affected by oblique flows of as much as 15^0.

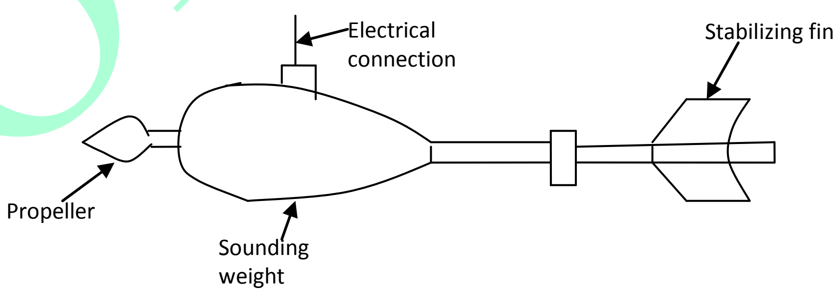


Fig.4.7: Horizontal axis current meter

Relationship between current meter rotation speed and stream velocity

A current meter is so designed that its speed of rotation varies linearly with stream velocity (V). The relationship is
$V = a N_s + b$
where V= stream velocity (m/s)
N_s = revolutions per second of current meter
a, b = constants

Calibration of current meter

Determination of constants a and b is known ascalibration of current meter.

Current meters are calibrated in ponds or long channels where water is held stationery. A vehicle with cantilever arm projection to the channel helps to lower and move the current meter in the pond water. For each run, the current meter is moved at a predetermined speed (v) and the number of revolutions of the meter (N_s) are counted. This experiment is repeated over a complete range of velocities and a best fit linear relation is developed.

Float, types, velocity rod

4.5 Discharge measurement by floats

Floats are used to measure velocity for a small stream in flood, small stream with rapidly changing water surface and for preliminary analysis.

$$V_s = \frac{L}{t}$$

V_s = surface velocity
L = Distance travelled
t = time taken to travel the float

Discharge = V_{av} x A
V_{av} = average velocity = 0.85 to 0.95 times surface velocity
A = cross-sectional area

Types of floats

- Surface float: a simple float moving on stream surface, wooden or metallic object, leaf, orange
- Subsurface float: two floats tied together by thin cord, one float submerged
- Rod float: cylindrical rod partly submerged

Method

- Select straight reach free from current and eddies.
- Measure distance between upstream and downstream section.
- Divide the cross-section into a number of subsections.

- For each subsection, release float at an upstream section and note the time taken by float to reach downstream section. Find average velocity of different sections and compute discharge.

4.6 Slope area method

This is indirect method of discharge measurement. In this method, Manning's equationand Bernoulli's equation are used to estimate the discharge for high floods based on previous flood marks. Two sections along a river reach are selected. The cross-sectional area of each section and the longitudinal profile between the sections is measured.

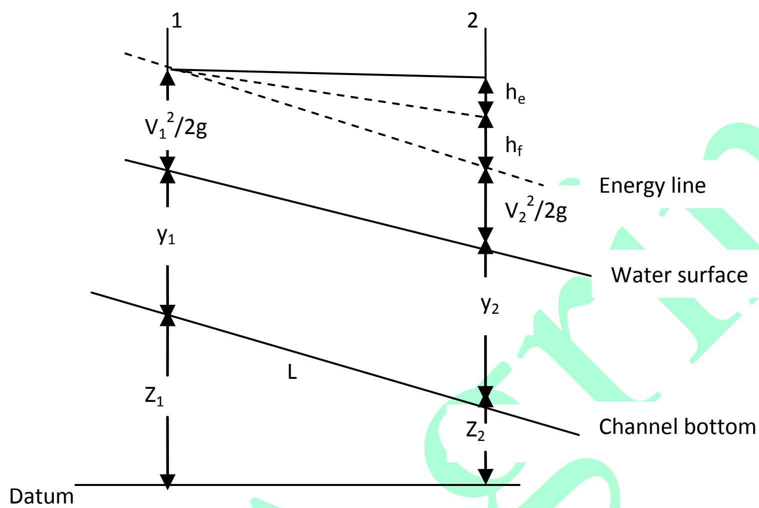


Fig. 4.8: Slope-area method

By using Bernoulli's equation for sections 1 and 2

$$Z_1 + y_1 + \frac{V_1{}^2}{2g} = Z_2 + y_2 + \frac{V_2{}^2}{2g} + h_L$$

Z_1, Z_2 : Datum headat sections1 and 2
y_1, y_2 : water depthat sections1 and 2
V_1, V_2: velocities at sections1 and 2
h_L = Head loss
$h_L = h_f + h_e$
h_f = Frictional loss
h_e = eddy loss

Denoting Z + y = h = water surface elevation above the datum

$h_1 + \frac{V_1{}^2}{2g} = h_2 + \frac{V_2{}^2}{2g} + h_f + h_e$

$h_f = (h_1 - h_2) + \left(\frac{V_1{}^2}{2g} - \frac{V_2{}^2}{2g}\right) - h_e$ (a)

From Manning's equation,

$Q = \frac{1}{n} A R^{2/3} S_f^{1/2}$

Q =Discharge
n = Roughness coefficient
A = Cross-sectional area
R = Hydraulic radius = A/P where P = wetted perimeter
S_f= Slope of energy line between two points
In other form, Manning's equation is expressed as
$Q = K\sqrt{S_f}$ (b)
K = Conveyance of channel = $\frac{AR^{2/3}}{n}$
$S_f = \frac{h_f}{L}$
For two sections, average conveyance is
$K = \sqrt{K1K2}$ (c)
Where $K1 = \frac{A1R1^{2/3}}{n1}$ and $K2 = \frac{A2R2^{2/3}}{n2}$
Eddy loss is given by
$h_e = K_e\left|\frac{V_1^2}{2g} - \frac{V_2^2}{2g}\right|$where K_e = Eddy loss coefficient

Procedure to compute peak discharge by using slope-area method

1. Compute cross-sectional area, wetted perimeter, hydraulic radius, K1and K2 at section 1 and 2. Compute K using $K = \sqrt{K1K2}$.
2. For first iteration, assume $V_1 = V_2$. This leads to $h_f = h_1 - h_2$ = Fall in water surface between two sections. So take $h_f = h_1 - h_2$.
3. Calculate Q using eq.$Q = K\sqrt{S_f} = K\sqrt{h_f/L}$.
4. Compute $V_1 (= Q/A_1)$ and $V_2 (= Q/A_2)$.
5. Now calculate a refined value of h_f by using eq.$(h_f)refined = (h_1 - h_2) + \left(\frac{V_1^2}{2g} - \frac{V_2^2}{2g}\right) - h_e$.
6. Take refined value of h_f for next iteration and repeat steps 3 to 5until the difference between two successive values of h_f is negligible.
7. Compute Q using final value of h_f.

Recommended criteria

- Distance between two sections = 75 times flood depth
- Fall in water head>15 cm
- Straight and uniform reach
- Quality of high-water marks should be good.

Difference between slope-area and velocity area method
I. Velocity-Area method is a direct method for discharge measurement, whereas slope area method is an indirect method of discharge measurement
II. In velocity-Area method, measurement is performed across a cross-section. The cross-section of a river is divided into a number of subsections by imaginary verticals. The depth of flow and velocity is measured at each vertical.
In slope area method, two sections along a river reach are selected. The cross-sectional area of each section and the longitudinal profile of high flood line between the sections is measured.
III. In velocity-Area method, velocity is measured by current meter or floats. In slope area method, velocity is computed from the concept of Hydraulics.
IV. In slope area method, the segmental discharge is obtained by multiplying segmental area and mean velocity, and the total discharge is obtained by summing the segmental discharge.

In slope area method, the computation of discharge is based on Manning's equation and Bernoulli's equation.
V. No trial and error is needed in velocity area method, whereas h_f is obtained by trial and error approach in slope area method.

4.7 Flow measuring structures

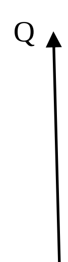
These structures produce unique control section in the flow. For such structures,
Q =f(H)
Q= discharge
H = Water surface elevation measured from a specified datum
Free flow: flow independent of downstream water level
Submerged or drowned flow: flow affected by downstream water level

Various structures
a. Thin plate structures: made of metal plate, e.g. V-notch, rectangular notch
b. Long base weirs (Broad-crested): made of concrete or masonry
c. Flume: Channel having constriction

Formula
Rectangular notch: $Q = \frac{2}{3}Cd\sqrt{2g}.LH^{1.5}$
V-notch: $Q = \frac{8}{15}C_d\sqrt{2g}Tan\frac{\theta}{2}H^{5/2}$
Broad-crested weir: $Q = C_d\,Lh\sqrt{2g(H-h)}$ and $Q_{max} = 1.705C_d\,LH^{3/2}$

4.8 Rating Curve (Stage-discharge relationship)

The relationship between discharge (Q) and stage (H) is known as rating curve. Continuous measurement of discharge is not feasible as it is costly and time consuming. So, discharge data with corresponding stage is collected from time to time as sample data, and a relation between stage and discharge (rating curve) is prepared from the sample data. As it is easy and inexpensive, stage is measured continuously. The rating curve is used to convert the measured stage into discharge. In this way, the continuous discharge value is obtained.



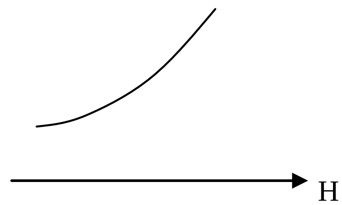


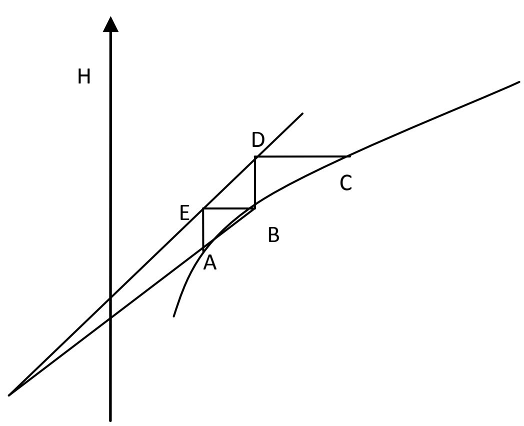
Fig. 4.9: Rating curve

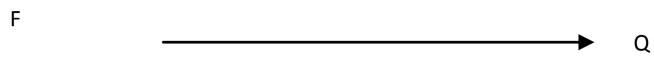
Equation of rating curve or stage-discharge relation is expressed in following exponential form.
$Q = a(H - H_0)^b$
Q = Discharge
a, b = Constants
H = Stage
H_0 = Stage for zero flow
Converting this equation to logarithmic form gives simple linear equation, which is then easy to use for further analysis.
$logQ = blog(H - H_0) + loga$
Or, Y=bX+c
Y =logQ, X = log(H-H_0)
The rating curve remains valid so long as the condition at a site remains stable.

Determination of parameters of rating curve
Determination of H_0
1. Graphical approach
a. Plot H and Q on a plain graph and draw best fit curve. Extrapolate the curve backwards to touch ordinate axis where Q =0. Take H corresponding to Q = 0 as H_0. Plot Q and H-Ho in log scale and check whether the plot is straight line. If not, take another value of H_0 close to it and repeat the procedure until straight line plot is obtained.
b. Plot Q vsH to an arithmetic scale and fit the smooth curve. Select three points on the curve such that their discharges are in geometric progression.
i.e.$Q2 = \sqrt{Q1Q3}$
Note H1, H2 and H3corresponding to Q1, Q2 and Q3. Compute H_0 by

$$H_0 = \frac{H1H3 - H2^2}{H1 + H3 - 2H2}$$

c. Plot Q vsH to an arithmetic scale and fit the smooth curve. Select three points A, B, and C on the curve such that their discharges are in geometric progression.Draw vertical lines at A and B and horizontal lines at B and C. Then two straight lines ED and BA are drawn to intersect at F as show in figure. The ordinate of F is required value of H_0.





2. Determine a, b and H_0 simultaneously by least square optimization method.

Value of a and b from regression

$$b = \frac{N(\sum XY) - (\sum X)(\sum Y)}{N(\sum X^2) - (\sum X)^2}$$

$$c = \frac{\sum Y - b(\sum X)}{N}$$

a $=10^c$

Coefficient of correlation

$$r = \frac{N(\sum XY) - (\sum X)(\sum Y)}{\sqrt{N(\sum X^2) - (\sum X)^2}\sqrt{N(\sum Y^2) - (\sum Y)^2}}$$

Methods for extension of rating curve

a. Extension based on Logarithmic plotting of rating curve or using the rating equation
b. Velocity area method: Extend stage-velocity and stage-area curve.
c. Conveyance slope method based on Manning equation:
Flood marks in the river course provides water surface slope of the peak.

$$Q = \frac{1}{n}AR^{2/3}S^{1/2} = K\sqrt{S}$$

K = Conveyance of channel = $\frac{AR^{2/3}}{n}$

- For different stages, compute K.
- Compute S for different stages by using $S = \frac{Q^2}{K^2}$ from available Q-h data.
- Plot and extend stage-conveyance(K) and stage-slope (S) curve. For different stage, take K and S from the two curves and compute Q as $K\sqrt{S}$.

Assumptions: For higher stages, the slope remains constant

<u>Control</u>

Control is combined effect of channel and flow parameters, which govern the stage-discharge relationship.

If the rating curve does not change with time, the control is called permanent control. In other words, the station with permanent control has single valued rating curve.

If the rating curve changes with time, it is called shifting control.

Shifting controls

- Vegetation growth, dredging or channel encroachment: no unique rating curve
- Aggradation or degradation in alluvial channel: no unique rating curve
- Variable backwater effect: same stage indicating different discharges
- Unsteady flow effects of rapidly changing stage: for the same stage, low discharge during rising and high discharge during falling (looped rating)

Correction for backwater effect: To take into account the backwater effect, secondary gauge is installed at some distance downstream of gauging site and the readings of both gauges are taken. Then, the fall of water surface in the reach is computed. The relationship for actual discharge (Q) is given by

$$\frac{Q}{Q_0} = \left(\frac{F}{F_0}\right)^m$$

Q_0 = Normalized discharge at the given stage when fall = F_0, when the stage in the river is same in both cases

F = Actual fall

m = exponent $\approx$0.5

Unsteady flow correction: Correction has to be applied in case of unsteady flow due to flood wave. The actual discharge (Q) under unsteady condition is given by

$$Q = Q_0\sqrt{1 + \frac{1}{V_w S_0}\frac{dh}{dt}}$$

Q_0 = discharge under steady flow conditions

V_w = Velocity of flood wave

S_0 = Bed slope of river

dh/dt = rate of change of stage

4.11 Estimating mean monthly flow for ungauged basin of Nepal

a. Medium irrigation project (MIP) Method

The MIP method presents a technique for estimating the distribution of monthly flows throughout a year for ungauged locations. For application to ungauged sites, it is necessary to obtain one flow measurement in the low flow period from November to April.

- In the MIP Method, Nepal has been divided hydrologically into seven zones. Once the catchment area of the scheme, one flow measurement in the low flow period and the hydrological zone is identified, long-term average monthly flows can be determined by multiplying the unit hydrograph (of the concerned region) with the measured catchment area.
- Hydrological zone can be identified based on the location of the scheme in the hydrologically zoned map of Nepal.
- For catchment areas less than 100 km^2, MIP method is used for better results.

If the measured date is on 15th of the particular month, the coefficient given in the table is directly used. For other date of measurement, coefficient for that date is found by interpolation.

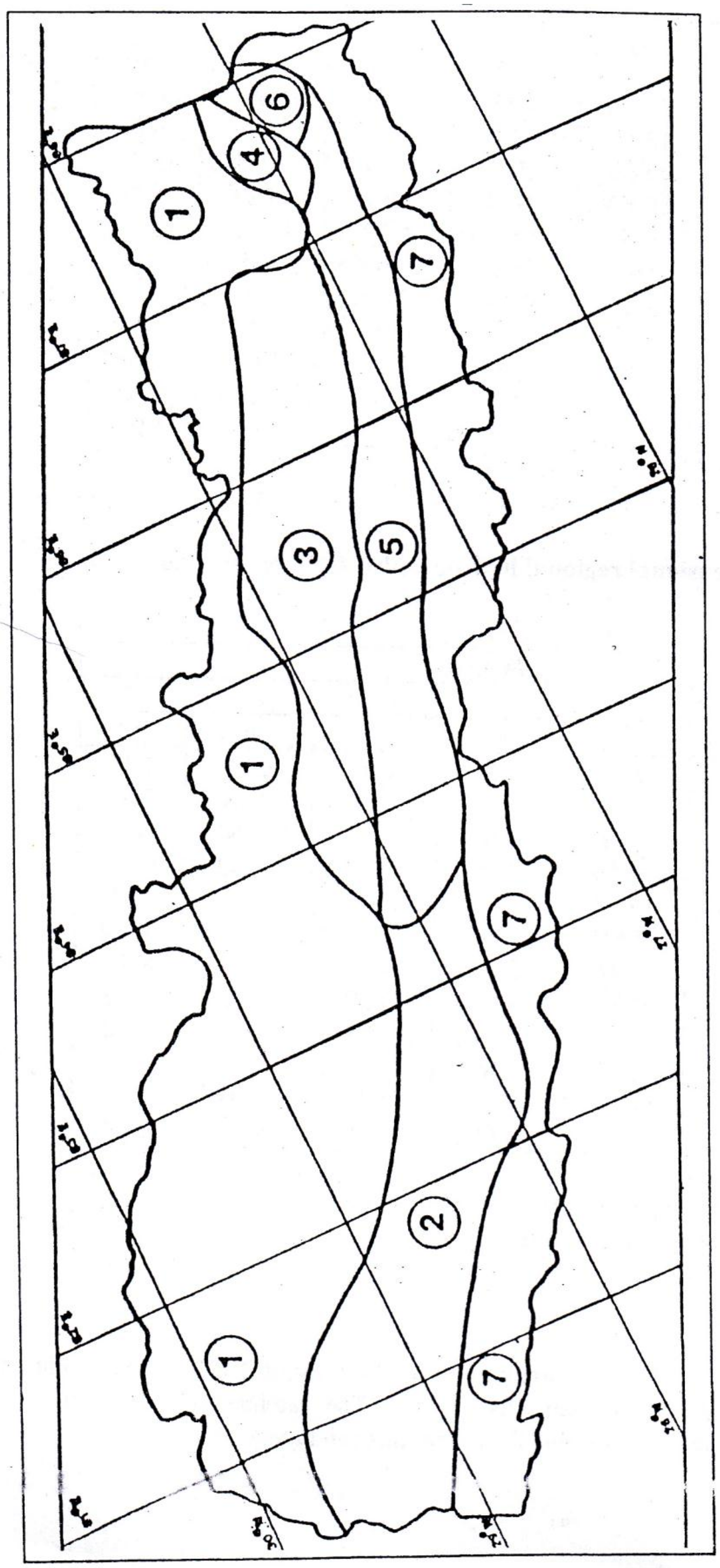
$$April\ flow = \frac{1}{coefficient\ of\ a\ particular\ month} Measured\ discharge$$

Monthly flow = April flow x Monthly coefficient

MIP non-dimensional regional hydrographs (Coefficient)

Month	Region						
	1	2	3	4	5	6	7
May	2.60	1.21	1.88	2.19	0.91	2.57	3.50
Jun	6.00	7.27	3.13	3.75	2.73	6.08	6.00
Jul	14.50	18.18	13.54	6.89	11.21	24.32	14.00
Aug	25.00	27.27	25.00	27.27	13.94	33.78	35.00
Sep	16.50	20.19	20.83	20.91	10.00	27.03	24.00
Oct	8.00	9.09	10.42	6.89	6.52	6.08	12.00
Nov	4.10	3.94	5.00	5.00	4.55	3.38	7.50
Dec	3.10	3.03	3.75	3.44	3.33	2.57	5.00
Jan	2.40	2.24	2.71	2.59	2.42	2.03	3.30
Feb	1.80	1.70	1.88	1.88	1.82	1.62	2.20
Mar	1.30	1.33	1.38	1.38	1.36	1.27	1.40
Apr	1.00	1.00	1.00	1.00	1.00	1.00	1.00

Note: all values for mid month



1. Mountain catchments.
2. Hills to the north of Mahabarat, river rising north of Siwaliks, inner terai.
3. Pokhara, Nuwakot, Kathmandu, Sunkosi tributaries.
4. Lower Tamur Valley
5. Rivers draining Mahabarats.
6. Kankai Mai basin.
7. Rivers draining from Churia range to the terai.

Hydrological regions of Nepal for MIP method

b. WECS/DHM (Hydest) Method

It is developed for predicting river flows for catchment areas larger than 100 km^2 of ungauged rivers based on hydrological theories, empirical equations and statistics. For long term average monthly flows, all areas below 5000m are assumed to contribute flows equally per km^2 area.

The average monthly flows can be calculated by the equation:

$$Q_{mean,(month)} = C \times (\text{Area of Basin})^{A1} \times (\text{Area below 5000m} + 1)^{A2} \times (\text{Mean Monsoon precipitation})^{A3}.$$

Where $Q_{mean(month)}$ is the mean flow for a particular month in m^3/s, C, A1, A2 and A3 are coefficients of the different months.

The catchment area can be calculated from the topographical maps (maps that show contours) once the intake location is identified.

The input data required in the equation are total basin area (km^2), basin area below 5000m (km^2) and the average monsoon precipitation (km^2) estimated from isohyetal map.

Values of coefficients for WECS/DHM method

Month	C	A1	A2	A3
Jan	0.01423	0	0.9777	0
Feb	0.01219	0	0.9766	0
Mar	0.009988	0	0.9948	0
Apr	0.007974	0	1.0435	0
May	0.008434	0	1.0898	0
Jun	0.006943	0.9968	0	0.2610
Jul	0.02123	0	1.0093	0.2523
Aug	0.02548	0	0.9963	0.2620
Sep	0.01677	0	0.9894	0.2878
Oct	0.009724	0	0.9880	0.2508
Nov	0.001760	0.9605	0	0.3910
Dec	0.001485	0.9536	0	0.3607

c. Catchment Area Ratio Method (CAR Method)

If the two catchments are hydrologically similar, then the extension of hydrological data for proposed site under study could be done simply by multiplying the available long term data at hydrologically similar catchments (HSC) with ratio of catchment areas of base (proposed site under study) and index (HSC) stations.

$$Q_b = Q_i \frac{A_b}{A_i}$$

Where, Q = discharge in m^3/s
A = drainage area in sq.km
Suffix 'b' stands for base station and i stands for index station.

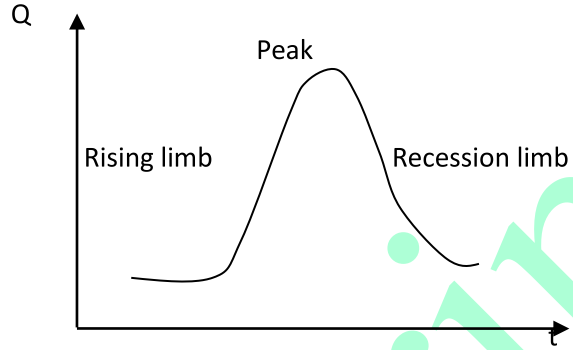
This method is useful if the hydro-meteorological data of the index station having similar catchment characteristics with the base station are available for the data extension.

Chapter 5: Hydrograph analysis

5.1 Hydrograph

Hydrograph is a graphical plot of discharge (Q) of a river at a given location over time.
It is the output or total response of a basin.

Components of hydrograph



Hydrograph

1. Rising limb
It is ascending portion of hydrograph. It is influenced by storm and basin characteristics. The rising limb rises slowly in the early stage of flood but more rapidly towards the end portion. This is because in the initial stage the losses are high. The flow begins to build up in the channel as the storm duration increases. It gradually reaches the peak when maximum area contributes.

2. Peak or crest segment
It is the part which contains peak flow, which is of interest to hydrologists. Peak of hydrograph occurs when all portions of basins contribute at the outlet simultaneously at the maximum rate. Depending upon the rainfall-basin characteristics, the peak may be sharp, flat or may have several well defined peaks.

3. Recession limb

Recession limb represents withdrawal of water from the storage built up in the basin during the earlier phase of the hydrograph. It extends from the point of inflection at the end of the crest to the beginning of natural groundwater flow. The recession limb is affected by basin characteristics only and independent of the storm.

Equation for recession curve

$$Q_t = Q_0 K_r^t$$

Q_0: initial discharge
Q_t: discharge at a time interval of t days
K_r: recession constant

Alternative form

$$Q_t = Q_0 e^{-at}$$

Where a = $-\ln K_r$

Terms

Time to peak: time lapse between starting of the rising limb to the peak

Time lag: time interval between centre of mass of rainfall hyetograph to the centre of mass of runoff hydrograph.

Time of concentration: time taken by a drop of water to travel from the remotest part to the outlet

Time base of hydrograph: time between starting of runoff hydrograph to the end of direct runoff due to storm.

5.2 Direct runoff and base flow

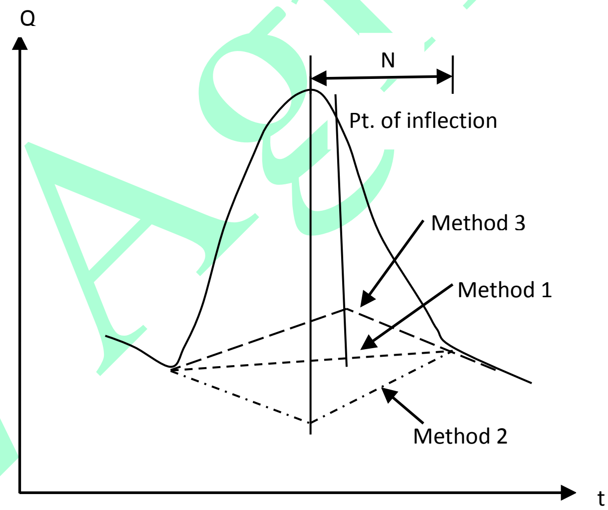
Direct runoff

It is the part of precipitation which appears quickly as flow in the river. (direct runoff = surface +subsurface)

Base flow

The part of runoff which receives water from the groundwater storage is called base flow.

Base flow separation



Base flow separation methods

1. Straight line method

Join the beginning of surface runoff to a point on the recession limb representing the end of direct runoff.

End point: by expert judgment or empirical equation

Empirical equation to find end of direct runoff

N = 0.83 $A^{0.2}$

N = time interval from the peak to the end of direct runoff

A = Basin area

2. Extend the base flow curve prior to the commencement of surface runoff till it intersects the ordinate drawn at the peak point. Join this point to the end point of direct runoff

3. Extend the base flow recession curve backwards after the depletion of flood water till it intersects the ordinate at the point of inflection. Join this point to the beginning of the surface runoff by smooth curve.

Direct runoff hydrograph: the surface runoff hydrograph obtained after separating base flow

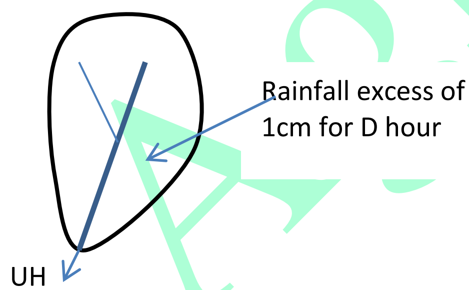
Types of stream

- Perennial: always carry flow
- Intermittent: limited contribution from groundwater
- Ephemeral: no base flow

Yield: total quantity of water that can be expected from a stream in a given period.

5.3 Unit Hydrograph

A unit hydrograph (UH) of a basin is defined as a direct runoff hydrograph (DRH) resulting from one unit depth of rainfall excess generated uniformly over the basin at a constant rate for an effective duration (D). The term unit refers to a unit depth of rainfall excess which is 1cm in SI unit and 1 inch in FPS unit. (Rainfall excess/effective rainfall = rainfall-loss)



Duration of unit hydrograph (D-hour UH): duration of rainfall excess
Assumptions

- Constant intensity of excess rainfall within the effective
- Uniform distribution of excess rainfall over the basin
- Constant base time of the DRH for excess rainfall of given duration
- Linear model: principle of superposition and proportionality holds
- Principle of time invariance holds
  -Given excess rainfall will always produce the same DRH whatever may be the season of the year (unchanging basin characteristics)

Principles applied in UH

I. Linearity principles
Linear relationship means output varies linearly with input. This principle is expressed by convolution theorem.

If $I(\tau)$ is intensity of input at time τ and $u(t-\tau)$ is the unit response after time t, then total response is given by $Q(t) = \int_0^t I(\tau)\, u(t-\tau) d\tau$. This is convolution integral.

There are two principles of linearity.
a. Principle of proportionality: If a solution y is multiplied by a constant c, the resulting function cy is also a solution.
r_e = excess rainfall, UH = Unit hydrograph (solution)
Output (DRH) = r_e*UH
b. Principle of superposition: If two solutions y1 and y2 of the equation are added, the resulting function y1+y2 is also a solution of the equation.
r_{e1}, r_{e1} = excess rainfall at t hr interval, UH = Unit hydrograph (solution)
Output (DRH) = (r_{e1}*UH)+ (r_{e2}*UH lagged by t hr)

II. Principle of time invariance: Given excess rainfall will always produce the same DRH whatever may be the season of the year (unchanging basin characteristics)

Features
- Rainfall excess (r_e) =1cm, runoff depth (r_d) = 1cm

  Continuity: Total depth of rainfall excess = total depth of direct runoff
- Runoff volume (V_d) = Basin area(A) x r_d = Ax1cm
- Rainfall intensity: 1/D in cm/h
- Lumped response: catchment as a single unit
- Initial loss absorbed by basin, no effect of antecedent storm condition

Applications of UH
- Computation of flood hydrograph for the design of hydraulic structures
- Extension of flow records at a site
- Flood forecasting
- Comparing the basin characteristics

Limitations of UH
- Minimum basin size> 2km^2, Maximum basin size up to 5000 km^2
- Not suitable for very long basins
- Applicable for short duration
- Not very suitable for basins having large snow cover
- UH is not applicable for basins having large storages
- UH is not applicable for basins having high variation of rainfall intensity.

5.4 Derivation of unit hydrograph

Selection criteria for flood hydrograph
- Selection of isolated storms occurring individually
- Fairly uniform rainfall over the entire basin
- Duration of rainfall: 1/5 to 1/3 of basin lag
- Range of rainfall excess: 1 to 4 cm

1. Derivation of UH for single storm
Given: streamflow data (Q) and basin area (A)
Single storm: all of the rainfall excess occurs at a reasonably uniform rate over a fairly short time period

- Separate baseflow (BF).
- DRH=Q-BF
- Volume of DRH (V_d) = $\sum DRH * \Delta t$
- CRunoff depth (r_d) = V_d/A
- UH= DRH/r_d

Effective duration of UH = Duration of excess rainfall.
Check whether total depth of runoff = total rainfall excess

2. Derivation of UH for multiple storms
Multiple storms: relatively long and varying intensities if rainfall
Storms: divided into number of equal periods and fairly constant rate of rainfall for each period
Duration of UH = Duration of period of each storm

De-convolution method
Given: DRH data and rainfall excess data
(If DRH is not given, compute base flow and compute DRH by subtracting baseflow from streamflow data)
Convolution Equation in discrete form
$Q_n = \sum_{m=1}^{n\leq m} P_m U_{n-m+1}$
n = number of runoff ordinates
m = number of periods of rainfall excess
Q_n = Direct runoff
P_m= Excess rainfall
U_{n-m+1} = UH ordinate

Use above equation for computing ordinate of UH with excess rainfall and direct runoff data.
For complex multi-peaked hydrograph: solution of above equation by least square regression.

5.5 Computation of runoff from given UH

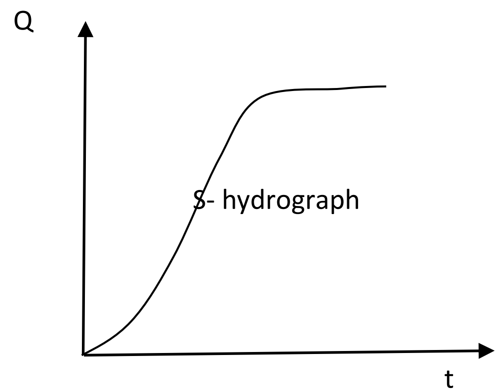
1. Single storm
- DRH =UH * rainfall excess.
- Total runoff=DRH+BF

2. Multiple storms
- Given: UH and effective rainfall for multiple durations

Use principle of proportionality and superposition
- DRH1=UH * first rainfall excess.
- DRH2=UH*second rainfall excess lagged by duration of first and second rainfall
  DRH3 = UH*third rainfall excess lagged by duration of first and third rainfall
  So on….
- DRH= DRH1+DHR2+DHR3+……
- Total runoff=DRH+BF

5.6 S-Hydrograph

S Hydrograph is a hydrograph resulting from a continuous excess rainfall at a constant rate of 1cm/h for an indefinite period. It is a theoretical concept. The curve is named S hydrograph as it looks like deformed S shape. The curve is obtained by adding a series of D-h unit hydrographs spaced at D-h apart.



The S-curve reaches a maximum equilibrium discharge at a time equal to the time base of the first unit hydrograph.
Unit rainfall excess = 1 cm in D hr
Rainfall intensity = 1/D in cm/hr
If A = basin area in km^2 and D is in hour, then
Equilibrium discharge $(Q_s) = \left(\frac{1}{D} x \frac{1}{3600x100}\right)(Ax10^6) = 2.778\frac{A}{D}$ m^3/s

Construction of S-curve
U(t) = S(t) – S(t-D)
S(t) = U(t)+ S(t-D)
where D = Duration of UH, S(t) = ordinate of S-curve at t, U(t) = ordinate of UH at t, S(t-D) = ordinate of S-curve at t-D

In other words,
Ordinate of S-curve at t = ordinate of D-hr UH at t + S-curve addition at time t
For $t \leq D$, S(t-D) = 0.

5.7 Computation of Unit hydrograph of different durations

In the computation of flood hydrograph, if the duration (D) of given UH and the duration (D') of excess rainfall is different, then the UH of D hour should be converted to UH of D' hour.

Given: UH of duration D
To compute: UH of duration D'
n = D'/D
If n is integer, use superposition method or S-curve method.
If n is real, use S-curve method.

a. Superposition method

- Lag the UH ordinate by D, 2D, ….(n-1)D.
- U1=Sum of the ordinates of all UHs.
- Ordinate of D'-hour UH = U1/n

b. S Hydrograph method

- Compute S-curve addition (=S(t-D)).
- Compute the ordinate of S-curve.
  S1 = UH(t)+ S(t-D)
- Lag the ordinates of S1 hydrograph by the duration D'. This is S2.
- Ordinate of D'-hour UH= (S1-S2)/n

In case of D'<D and the time interval of data is not equal to D', first plot the given UH and read the values with time interval equal to D'. Then follow above steps.

If the ordinates of UH becomes negative or shows fluctuations in the tail part, then manually smoothen the tail part.

Basic Numericals of Unit Hydrograph

<u>Derivation of UH</u>

Single storm

Given below are the observed flows from a storm of 4hr duration on a stream with a catchment area of 613 km^2. Derive 4hr unit hydrograph. Make suitable assumptions regarding base flow.

Time (hr)	0	4	8	12	16	20	24	28	32	36	40	44	48
Observed flow (m3/s)	10	110	225	180	130	100	70	60	50	35	25	15	10

Solution:
Catchment area (A) = 613 km^2
Assume base flow (BF) = 10 m^3/s
Direct runoff (Q_{dr}) = Q-BF
Volume of runoff (V) = $\sum Q_{dr}\,\Delta t$
Runoff depth (r_d) = V/A
Divide Q_{dr} by r_d to get UH ordinate.
Δt is same for each runoff ordinate.
Δt = 4 hour = 4x3600 s
V = $\sum Q_{dr}\,\Delta t = \Delta t \sum Q_{dr}$ = 890x4x3600
$r_d = \frac{V}{A} = \frac{890\times4\times3600}{613\times10^6}$= 0.02m =2cm

Time (hr)	0	4	8	12	16	20	24	28	32	36	40	44	48
Q (m^3/s)	10	110	225	180	130	100	70	60	50	35	25	15	10
BF (m^3/s)	10	10	10	10	10	10	10	10	10	10	10	10	10
Qdr (m^3/s)	0	100	215	170	120	90	60	50	40	25	15	5	0
UH (m^3/s)	0	50	108	85	60	45	30	25	20	13	7.5	2.5	0

The ordinates of a hydrograph of a surface runoff (DRH) resulting from 4.5cm of rainfall excess of duration 8hr in a catchment are as follows:

Time (hr)	0	5	13	21	28	32	35	41	45	55	61	91	98	115	138
Discharge (m^3/s)	0	40	210	400	600	820	1150	1440	1510	1420	1190	650	520	290	0

Derive the ordinates of 8hr-unit hydrograph.

Solution:
Direct runoff = Q
Rainfall excess (R_e) = 4.5cm
For single storm, UH ordinate = Q/R_e

Time (hr)	0	5	13	21	28	32	35	41	45	55	61	91	98	115	138

Discharge (m^3/s)	0	40	210	400	600	820	1150	1440	1510	1420	1190	650	520	290	0
UH	0	8.9	46.7	88.9	133.3	182.2	256	320	335.6	316	264.4	144	116	64.44	0

Multiple storm

The following table gives the ordinates of a DRH resulting from two successive 3-hour durations of rainfall excess value of 2 cm and 4 cm respectively.

t (hr)	0	3	6	9	12	15	18	21	24	27	30
DRH (m^3/s)	0	120	480	660	460	260	160	100	50	20	0

Derive the ordinates of 3-hr UH.

Solution:
a. Effective rainfall, R_1 = 2 cm and R_2 = 4cm
It is a case of multiple storms. We have to use discrete time convolution equation to compute UH ordinate. The equation is

$$Q_n = \sum_{m=1}^{n \le m} R_m U_{n-m+1}$$

Q = Direct runoff, R = Excess rainfall, U = UH ordinate
Here, total no. of runoff ordinates (n) = 9
Total number of rainfall excess values (m) =2

For n =1, m =1
$Q_1 = R_1U_1$
$U_1 = Q_1/R_1$ =120/2 = 60
For n = 2, m=1, 2
$Q_2 = R_1U_2+R_2U_1$
$U_2 = (Q_2-R_2U_1)/R_1$ = (480-4x60)/2 = 120
For n = 3 onwards, m= 1, 2. So, we can use the similar expression as that of U_2 for n = 3 onwards.
$U_n = (Q_n-R_2U_{n-1})/R_1$

$U_3 = (Q_3-R_2U_2)/R_1$ = (660-4x120)/2 = 90
$U_4 = (Q_4-R_2U_3)/R_1$ = (460-4x90)/2 = 50
$U_5 = (Q_5-R_2U_4)/R_1$ = (260-4x50)/2 = 30
$U_6 = (Q_6-R_2U_5)/R_1$ = (160-4x30)/2 = 20
$U_7 = (Q_7-R_2U_6)/R_1$ = (100-4x20)/2 = 10
$U_8 = (Q_8-R_2U_7)/R_1$ = (50-4x10)/2 = 5
$U_9 = (Q_9-R_2U_8)/R_1$ = (20-4x5)/2 = 0

Resulting UH

t (hr)	0	3	6	9	12	15	18	21	24	27	30
UH(m^3/s)	0	60	120	90	50	30	20	10	5	0	0

UH to flood hydrograph

The ordinate of a 4-h UH of a catchment of area 1000km^2 are given below. Calculate flood hydrograph resulting from two successive 4-h storms having rainfall of 1.5cm each. Assume uniform base flow of 10 m^3/s and ϕ-index equal to 0.10 cm/hr.

t(hr)	0	4	8	12	16	20	24	28	32	36	40	44
4hr UH (m^3/s)	0	20	60	150	120	90	66	50	32	20	10	0

Solution:

ϕ-index (infiltration loss) = 0.1 cm/hr

For 4 hour, loss (L) = 4x0.1 =0.4 cm

Rainfall values, R1 = 1.5 cm and R2 = 1.5 cm

Rainfall excess (R_{e1}) = R1-L = 1.5-0.4 =1.1cm

Rainfall excess (R_{e2}) = R2-L = 1.5-0.4 =1.1cm

DHR1 = UHxR_{e1}

DRH2 =UHxR_{e2} (lagged by 4 hour)

DRH = DRH1+DRH2

Q = DRH + BF

Computation of flood hydrograph

t(h)	4 hr UH (m^3/s)	DRH1 (m^3/s)	DRH2 (m^3/s)	DRH (m^3/s)	BF (m^3/s)	Q (m^3/s)
0	0	0		0	10	10
4	20	22	0	22	10	32
8	60	66	22	88	10	98
12	150	165	66	231	10	241
16	120	132	165	297	10	307
20	90	99	132	231	10	241
24	66	72.6	99	171.6	10	181.6
28	50	55	72.6	127.6	10	137.6
32	32	35.2	55	90.2	10	100.2
36	20	22	35.2	57.2	10	67.2
40	10	11	22	33	10	43
44	0	0	11	11	10	21
(48)			0	0	10	10

UH of different durations

The ordinates of a 4 hour UH of a basin of area 25 km^2 are given below.

t (hr)	0	4	8	12	16	20	24	28	32	36	40	44	48	52
UH(m^3/s)	0	30	55	90	130	170	180	160	110	60	35	20	8	0

Calculate the following.
a) 4-hr DRH for a rainfall of 3.25cm with ϕ-index of 0.25cm.
b) a 12-hr UH by using the method of superposition.
c) a 12-hr UH by using the S-curve method.

Solution:
a) Rainfall (R) = 3.25 cm, ϕ-index = 0.25cm
Rainfall excess (re) = 3.25-0.25 = 3cm
DRH = UH xre

Computation of DRH

t (hr)	0	4	8	12	16	20	24	28	32	36	40	44	48	52
UH(m^3/s)	0	30	55	90	130	170	180	160	110	60	35	20	8	0
DRH (m^3/s)	0	90	165	270	390	510	540	480	330	180	105	60	24	0

b) Required duration of UH (D') =12 hr
Given duration (D) =4 hr
n = D'/D = 3 (integer)

UHa = UH lagged by 4 hr, UHb = UH lagged by 8 hour
UH1 = UH + UHa+ UHb
12hr-UH = UH1/(D'/D) = UH1/3

Computation of 12-hr UH using method of superposition

t (hr)	UH	Uha	Uhb	UH1	12-hr UH (m^3/s)
0	0			0	0
4	30	0		30	10
8	55	30	0	85	28.3
12	90	55	30	175	58.3
16	130	90	55	275	91.7
20	170	130	90	390	130
24	180	170	130	480	160
28	160	180	170	510	170
32	110	160	180	450	150
36	60	110	160	330	110
40	35	60	110	205	68.3
44	20	35	60	115	38.3
48	8	20	35	63	21
52	0	8	20	28	9.3

56		0	8	8	2.7
60			0	0	0

c. S-curve addition = Ordinate of S-curve at (t-D)
Ordinate of S curve (S1) = ordinate of UH+ S-curve addition
S 2 = S1 lagged by 12 hr hour
12hr-UH = (S1-S2)/(D'/D) = (S1-S2)/3

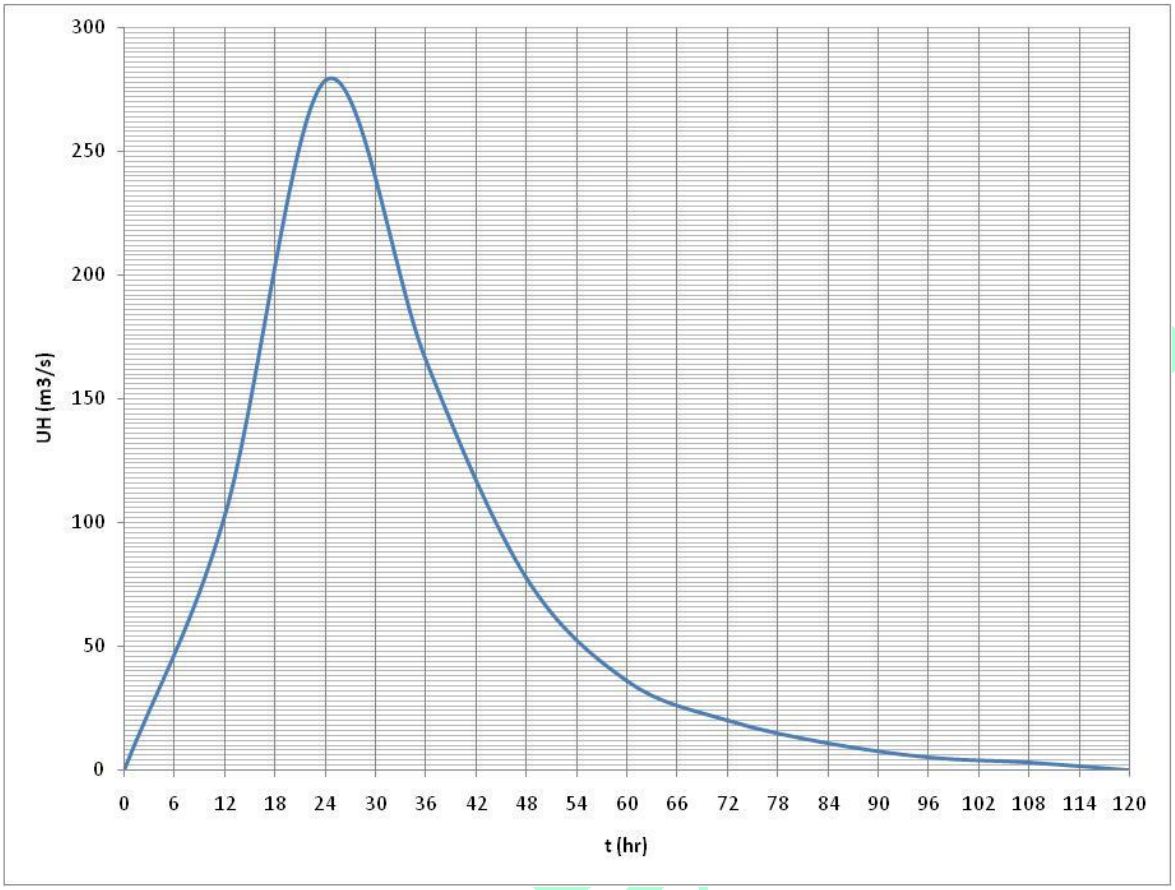
Computation of 12-hr UH using S-Curve method

t (hr)	UH	S-Curve addition	S-curve (S1)	S2	12-hr UH (m^3/s)
0	0		0		0
4	30	0	30		10
8	55	30	85		28.3
12	90	85	175	0	58.3
16	130	175	305	30	91.7
20	170	305	475	85	130
24	180	475	655	175	160
28	160	655	815	305	170
32	110	815	925	475	150
36	60	925	985	655	110
40	35	985	1020	815	68.3
44	20	1020	1040	925	38.3
48	8	1040	1048	985	21
52	0	1048	1048	1020	9.3
56			1048	1040	2.7
60			1048	1048	0
64			1048	1048	0

Given below is a 12-hr UH. Derive 6-hr UH.

t (hr)	0	12	24	36	48	60	72	84	96	108	120
UH(m^3/s)	0	103	279	165	78	36	20	11	5	3	0

Solution:
Required duration of UH (D') =6 hr
Given duration (D) =12 hr
n = D'/D = 0.5 (real)
Here, D'<D. To derive UH of 6 hr, the interval of ordinates of given UH should be at least 6hour.
Plot given UH versus t on a graph paper and get the values of UH at 6 hour interval.



S-curve addition = Ordinate of S-curve at (t-D)
Ordinate of S curve (S1) = ordinate of UH+ S-curve addition
S2 = S1 lagged by 6 hour
6-hr UH = (S1-S2)/(D'/D) = (S1-S2)/0.5

Computation of 6-hr UH

t (hr)	UH(m3/s)	S curve addition	S1	S2	6-hr UH	6-hr UH (corrected)
0	0		0		0	0
6	48		48	0	96	96
12	103	0	103	48	110	110
18	191	48	239	103	272	272
24	279	103	382	239	286	286
30	238	239	477	382	190	190
36	165	382	547	477	140	140
42	117	477	594	547	94	94
48	78	547	625	594	62	62
54	53	594	647	625	44	44
60	36	625	661	647	28	28

66	27	647	674	661	26	26
72	20	661	681	674	14	14
78	15	674	689	681	16	10
84	11	681	692	689	6	6
90	8	689	697	692	10	4
96	5	692	697	697	0	0
102	4	697	701	697	8	0
108	3	697	700	701	-2	0
114	2	701	703	700	6	0
120	0	700	700	703	-6	0
126		703	700	700	0	0

The UH of 6 hour should be corrected manually from 90 hour onwards to make it smooth.